A glacially induced incised valley system in the Late Ordovician (Hirnantian) of the Oslo Region. 

Sedimentary, sequence stratigraphic and carbon isotope analysis.

Franziska Franeck
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GAIUS IVLIVS CAESAR
Abstract

The Hirnantian Langøyene Formation in the central Oslo Region consists of mixed carbonate and silicilastic dominated sedimentary facies, which were deposited in a shallow, epicontinental sea on the Baltic craton. Sedimentological investigations were carried out at the localities on the southern shorelines of Langøyene, Rambergøya and Hovedøya in the inner part of the Oslofjorden. Sedimentary logs have been drawn from two sections on Langøyene and Rambergøya, respectively, and nine sections on Hovedøya. Two major and one minor unconformities have been identified within the Langøyene Formation and interpreted to have formed due to sea-level changes brought about by the Hirnantian glaciation in Gondwana. Observations from the northern shore of Hovedøya, Bleikøya and Gressholmen have also been included in the interpretation of the depositional system, presented as an incised valley model. A sea level curve, reconstructed on basis of the reference section on western Rambergøya, fits remarkably well to Hirnantian sea-level curves from Baltica, Laurentia and Gondwana. Evidence for glacioeustatic sea-level changes and accompanied erosional gaps in sedimentary successions has been widely recorded from Baltoscandia and Laurentia. Carbon isotope measurements show a significant Hirnantian isotope curve excursion (HICE) during the Latest Ordovician throughout the World. By the use of new carbon isotope data, the HICE is identified in the Husbergøya and Langøyene formations. The carbon isotope curve is compared with previously published isotope curves from Baltica and Laurentia.
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1 Introduction

Glaciations are commonly recorded in sediments through tillites and moraines close to the ice-centre (e.g. Hambrey, 1985; Eyles, 1993). In areas more distant to the actual ice-covered regions, one has to look for drop stones in marine sediments and indicators of glacially influenced changes, e.g. a fall in sea level or associated faunal changes to correlate them to the actual glaciation-related sediments (Berry and Boucot, 1973; Eyles, 1993). Evidence for a Late Ordovician (Hirnantian) glaciation (Hambrey, 1985) has been recorded in both, stratigraphic, lithological and stable isotope records, though its duration and extends led to controversial discussions (Marshall and Middleton, 1990; Brenchley et al., 1994; Delabroye and Vecoli, 2010; Loi et al., 2010). Challenges for the correlation of glacial-influenced strata are the determination of their stratigraphic age and tectonic stability of the platform or basin where sediments were deposited (McKerrow, 1979). Carbon isotope chemostratigraphy has become a frequently used tool for correlation, but this method also leads to some problems regarding the understanding of the actual isotope fractionation (Delabroye and Vecoli, 2010).

Epicontinental seas are very sensitive to sea level fluctuations due to their shallow depth (Wellner and Bartek, 2003; Nichols, 2009; Midtkandal and Nystuen, 2009; Glørstad-Clark et al., 2010, 2011). Hence, shallow seas may be suitable for detecting changes in relative sea level, which may have been triggered by the advances and retreat of continental ice sheets, and thereby glacioeustacy. Given the presence of fluvial drainage systems in the coastal area, rivers will most likely form incised valleys during a sea-level lowstand (Boyd et al., 1992). These incisions form isolated depositional environments that are subsequently filled with sediments during sea-level rise (Dalrymple, 2006). Sedimentary successions of incised valleys may provide information about dominant sedimentation processes and environment in the transgressive stage after the sea-level lowstand.

The Upper Ordovician Langøyene Formation in the Oslo Region is characterized by storm-dominated, open shelf sediments, formed in an epicontinental sea (Brenchley and Newall, 1975; Brenchley et al., 1979). A clearly regressive trend during the Ordovician, and the occurrence of at least one conglomerate representing the first infill above an erosional surface, are mentioned already in early descriptions (e.g. Spjeldnaes, 1957, 1961; Brenchley and Newall, 1975, 1977, 1980; Brenchley et al., 1979). A
reason for the uncertainties in interpreting the depositional environments may be the high lateral sedimentary facies variation (Brenchley and Newall, 1980). Despite the distance of the Oslo Region to the center of the continental Hirnantian glaciation on the Gondwana continent, Bjørlykke (1974a) and Brenchley and Newall (1980) suggested that glacioeustatic sea-level changes may be detectable in the Upper Ordovician of the Oslo Region and therefore be an important chronostratigraphic tool for the Ordovician of the region.

The Master Theses of Kjærsgaard (2014) and Sandbakken (2014) presented further data on stratigraphy and sedimentology from the Upper Ordovician on the islands Hovedøya, Rambergøya and Langøyene in the inner Oslofjorden and also forwarded the hypothesis of glacioeustatic control on the unconformities and conglomeratic strata in the upper part of the Langøyene Formation. The present study builds on their work as well as previously published material from the area.

It has also been suggested that the unconformities and associated conglomerate beds in the upper part of the Langøyene Formation may have been caused by tectonic uplifts, related to the Caledonian orogeny (e.g. Bjørlykke, 1983; Baarli, 1990) The Caledonian orogeny started during the Middle to Late Ordovician, proceeding into the Silurian, and also deformed the Early Palaeozoic strata by thrusting and folding in the Oslo Region, including the area of the present study (Nystuen, 1981; Bjørlykke, 1983; Morley, 1986, 1987; Bruton et al., 2010). Effects of this orogeny, accompanied by high tectonic activity connected to erosion of thrust sheets advancing from nowadays northwest to southeast, may have had a significant influence on the development of the depositional environment of the shallow marine sea of the Oslo Region during Ordovician time (Spjeldnæs, 1957; Bjørlykke, 1974a, 1983; Baarli, 1990).

The primary scope of the present work is to make a model for the development of the recorded unconformities and associated overlying sedimentary successions thorough a detailed sedimentary characterization and reconstruction of the sedimentary development in the Upper Ordovician Langøyene Formation. The study has been performed on strata exposed in the islands Hovedøya, Rambergøya and Langøyene in the inner part of the Oslofjorden. The working hypothesis is whether the glaciation on Gondwana is recorded in the Langøyene Formation. Furthermore, the scope of the study also includes whether the Hirnantian glaciation can be recorded as a single or multi-phased event in the strata of the Oslo Region, and to discuss the mechanisms relative to influence of Caledonian tectonic activity as cause of the incision in the Langøyene Formation. Carbon isotope values may detect the Hirnantian isotope curve excursion.
(HICE) in the Langøyene Formation. This may support the potential for correlation of the Upper Ordovician succession in the Oslo Region with Baltoscandian and global equivalents, and by this reason analysis of the stable carbon isotopes have been carried out as part of this study.
2 Palaeoclimatic framework

2.1. The Hirnantian glaciation

The Late Ordovician (Hirnantian) glaciation has drawn scientific interest not only because it is associated with one of the major mass extinctions (first of the “big five”), but also because of its appearance during a warm period in Earth’s climate (Sheehan, 2001; Herrmann et al., 2004b).

An early study about the climate during the Ordovician was presented by Spjeldnaes (1961), who tried to give a conclusive overview, but has to be viewed critically. Jaanusson (1973) suggested based on investigations of carbonates in the Baltoscandian epicontinental sea that deposition took place under subtropical to tropical conditions. Due to sea-level changes are the successions from the former epicontinental basin barely complete in any of the localities throughout Baltoscandia (Jaanusson, 1973), which may have been caused by glacioeustatic sea-level changes. However, tillites from the western Sahara (earlier Gondwana, cf. Fig. 3.2) gave first evidence of an extensive glaciation in the Late Ordovician (Hambrey, 1985). The ice sheet covered a considerable area of western Gondwana, including much of the North African and South American part (Scotese et al., 1999). The total affected area was estimated to $6 \times 10^6$ km$^2$ by Hambrey (1985) and $11 \times 10^6$ km$^2$ by Crowley and Baum (1991).

The duration of the Hirnantian glaciation became a subject of controversial discussions, based amongst others on evidence from $\delta^{13}$C isotopes (Saltzman and Young, 2005) or eccentricity cycles (Sutcliffe et al., 2000). Early estimations by Hambrey (1985) suggested durations of 35 Ma. Results of later studies showed evidence for a rather short-lived event, lasting for about 0,5 to 1 Ma (Brenchley et al., 1994, 2003) or even less (0,2 Ma) (Sutcliffe et al., 2000). As discussion continued, a start of the glaciation about 10 Ma before the Late Ordovician glacial maximum was suggested by Saltzman and Young (2005). However, the glacial maximum is set to the end of the Hirnantian (Saltzman and Young, 2005). This timing of glaciation will thus be relevant to the present study.

Studies by Herrmann et al. (2004a,b) focussed on possible reasons for the glaciation, based on atmospheric $p$CO$_2$. Their models show that there must have been an ini-
tially relative low sea-level to cause a glaciation during the relative warm climate at that time. Both a drop in sea-level at low atmospheric $p\text{CO}_2$ in the Late Ordovician and a drop in sea-level before the Late Ordovician would have given positive feedback on a glaciation during that time, since more exposed areas would have given space for ice sheets to grow (Herrmann et al., 2004a,b). The glaciation may not have happened without other environmental changes during that time, e.g. continental plate movements (southward movement of the Gondwana continent) and the accompanied sea-level drop (Herrmann et al., 2004a).

Two major glacial phases were described considering the Late Ordovician glacial maximum, showing at least two subdivisions, respectively (e.g. Sutcliffe et al., 2000; Ghienne, 2003; Ghienne et al., 2007). The two smaller order glacial cycles were reported from Mauritania, West Africa (Ghienne, 2003) and Jordan (Armstrong et al., 2009), as examples of areas close to the ice centre, and less well developed in Morocco and Turkey (Ghienne et al., 2007) from areas having been marginal to the center of glaciation (Delabroye and Vecoli, 2010). The second glacial phase seems to be the most widespread and significant one (Ghienne, 2003), and therefore may have had the strongest global significance in terms of glacioeustatic changes. Finnegan et al. (2011) found evidence for a connection of a global cooling of ca. $5^\circ\text{C}$, turnover in the carbon cycle and the Late Ordovician mass extinction.

2.2. Eustatic sea-level changes during the Late Ordovician and Early Silurian

The eustatic sea-level in the Early Palaeozoic shows a gradual increase and transgression from the Cambrian to the Early Ordovician, decreasing sea-level in the Middle Ordovician and then a substantial increase until the Early Katian, where the eustatic sea-level reached its maximum (McKerrow, 1979; Haq and Schutter, 2008). The sea-level at its maximum is estimated to have been about 225 m higher than present day sea-level (Haq and Schutter, 2008). During the Late Ordovician (Late Katian and Hirnantian), a short, regressive pulse occurred, most likely as a response to the Hirnantian glaciation (McKerrow, 1979; Haq and Schutter, 2008). This abrupt, but prominent shallowing (Haq and Schutter, 2008) can be observed at several marine platforms which locally may have become emergent (Berry and Boucot, 1973). A drop in sea-level of 80 to 130 m was suggested for Central Sweden (Baltica) (Kröger et al., 2015) and > 50 m for East Canada (Laurentia) (Desrochers et al., 2010). A subsequent transgression, starting
in the Early Silurian and reaching its maximum in the Middle Silurian, was followed by a decrease in sea-level until the Devonian (McKerrow, 1979; Haq and Schutter, 2008, and figures therein).

The relative sea-level curve for the Oslo Region by Nielsen (2004), in Fig. 3.4, shows a fluctuating, regressive trend in the Late Ordovician, followed by an Early Silurian transgression. Spjeldnæs (1957) suggested that this transgression started in the south of the Oslo Region, flooding previously emergent areas.

2.3. Climate zones and accompanying weather and current conditions

The Oslo Region was situated at low latitudes at the southern hemisphere, in the zone of trade winds, during Late Ordovician and Early Silurian times (Brenchley et al., 1979; Cocks and Torsvik, 2002, 2006) (Fig. 3.2). The trade winds are highly influenced by the Coriolis effect, and therefore deflected towards the west (Baarli, 1985). Studies from Eastern Canada (Ellis Bay Formation, Anticosti Island) showed that climatic conditions turned out to be more arid in low latitudes during the Hirnantian (Desrochers et al., 2010). Waves generated by wind have been interpreted to have had a high significance for the deposition of sediments in the shallow epicontinental seas in the Oslo Region during the Ordovician (Bjørlykke, 1974b). According to Bjørlykke (1974b) the winds of the Ordovician epicontinental sea created higher waves in areas of large depths, whereas the wave energy diminished in shallower parts of the sea, due to friction with the sea floor. Waves that were created in shallow parts of the sea would have less energy and therefore may have influenced less on the sedimentation.

Several authors suggested a storm origin of most of the sand layers in the Upper Ordovician and Lower Silurian succession in the epicontinental sea covering Oslo Region at that time (Brenchley et al., 1979; Baarli, 1985). This interpretation would fit the expected climatic conditions in these areas, influenced by strong winds, as they are mentioned above. Brenchley et al. (1979) interpreted the storm events to have been relatively unusual events, occurring approximately every 10 to 15 000 years, increasing to once every 5 to 10 000 years in the Silurian (Baarli, 1985). The devastating storms will rework sediments and therefore be the most important events in these latitudes to be preserved in the sedimentary record (Baarli, 1985).

Kjærsgaard (2014) found dispersed sand grains in the micritic Upper Ordovician limestone beds and concluded that also most of the micritic limestone units had been deposited during storms as mixed siliciclastic-carbonate mud suspensions.
3 Regional geology and stratigraphy

The Oslo Region has been the subject of numerous geological studies and scientific investigations through more than 200 years. The area was internationally acknowledged as a province of particular significance for the geological sciences through the famous travel report published in 1810 by the German geologist Leopold von Buch (Larsen et al., 2008). Systematic stratigraphic studies and geological mapping started with a series of publications by e.g. Kjerulf (1855, 1862). These pioneer studies were continued by Brøgger and Kjær, who mapped and described considerable parts of the region, including stratigraphical and palaeontological aspects (e.g. Brøgger, 1882, 1887; Kjær, 1897, 1908). Their works were carried out in great detail and have been of fundamental importance until the middle of the 20th century when new studies supplied additional data to the knowledge of the Oslo Region.

During the last 50 to 60 years, the main aspects of the geology in the Oslo Region have been re-described. A substantial part of the work on the Upper Ordovician, Lower Silurian successions was carried out by e.g. Spjeldnæs (1957); Brenchley and Newall (1975); Worsley et al. (1983); Baarli (1985) and Owen et al. (1990). For further information on previous studies in the Oslo Region, see Bockelie and Nystuen (1985) and Bruton et al. (2010).

The Oslo Region covers an area 115 km north and south of the city of Oslo and varies in width from 40 to 70 km (Bruton et al., 2010) (Fig. 3.1). The focus of this work concentrates on three of the islands in the inner part of the Oslo fjord: Hovedøya, Rambergøya and Langøyene. Additionally, localities on Gressholmen were considered (Fig. 3.1). Due to north-east, south-west trending folds in this area, the outcrops are aligned on the southern- and northernmost flanks of the mentioned islands (Brenchley and Cocks, 1982).

During the Early Palaeozoic, in the Cambrian and Ordovician, the Oslo Region was situated in a shallow epicontinental sea on the Baltic Shield (Brenchley et al., 1979; Bjørlykke, 1983; Baarli, 1985), at low latitudes on the southern hemisphere (Cocks and Torsvik, 2002, 2006) (Fig. 3.2). Whereas this area was subaerially exposed during the Precambrian, the Early Cambrian transgression caused flooding of the Oslo Region (Bjørlykke, 1974a; Worsley and Nakrem, 2008).
The Cambrian transgression formed a thin layer of conglomerate on top of the Sub-Cambrian Peneplane (SCP), followed by black alum shale (Worsley and Nakrem, 2008). The Upper Cambrian and Lower to Middle Ordovician succession shows a development from mud deposits to more calcareous-dominated sediments with an increasing amount of incoming sand (Brenchley and Newall, 1975; Bruton et al., 2010). This change of sedimentation character furthermore indicates the transition from an epicontinental basin towards a foreland basin during the development of the Caledonian orogen (Bruton et al., 2010).

With proceeding convergence of Baltica and Laurentia, closing the Iapetus Ocean that since Late Ediacaran times had separated these two continental plates, the regional variations of sedimentary facies in the Oslo Region and the Baltic continent increased (Worsley and Nakrem, 2008). Regional variations can be correlated with stable platform conditions to the east and the developing Caledonian orogen to the west (Bruton et al., 2010) (Fig. 3.3). This is also reflected in correlations of the Upper Ordovician...
successions in Scandinavia, showing relatively thin successions in Sweden and a successive deepening towards the Oslo Region (Bjørlykke, 1974a; Bruton et al., 2010). Bergström (1980) did studies on conodonts from the Oslo Region, which indicated a metamorphic alteration of the sediments at temperatures of about 300°C. A possible reason for this might be an increase in the geothermal flux and the magmatic activity during the Permian, generating plutons, lavas, dykes and sills (Bergström, 1980; Bruton et al., 2010).

3.1. Basin development in the Oslo Region

3.1.1. Cambrian to Late Ordovician epicontinental basin

Due to the Middle Cambrian transgression in the central and southern part of the Oslo Region, the Alum Shale Formation can nowadays be found in the lowermost succession of the flooded areas (Worsley and Nakrem, 2008). The Alum Shale is rich in or-
CHAPTER 3. REGIONAL GEOLOGY AND STRATIGRAPHY

Figure 3.3.: Platform conditions during the Late Ordovician and Early Silurian with the Caledonian thrust sheet coming from the East and leading to successive tectonic subsidence in former epicontinental sea covering the Oslo Region.

ganic matter, and therefore the bottom conditions of this ancient epicontinental sea must have been anoxic (Worsley and Nakrem, 2008). During the Ordovician, shales were deposited under more oxic conditions, since water circulation increased, and therefore these strata contain a rich fossil fauna (Worsley and Nakrem, 2008).

The tectonic conditions through the Cambrian and Ordovician were described by Bjørlykke (1983) to have been relatively stable. Though, changes in relative sea-level may have affected the character of sedimentation that mainly gave rise to dark grey to black shales, interbedded with carbonates and also some sandstone beds in the Upper Ordovician, showing a successive increase of both calcareous and siliciclastic components upwards (Bjørlykke, 1983; Owen et al., 1990; Worsley and Nakrem, 2008; Bruton et al., 2010) (Fig. 3.4). The end stage of the epicontinental basin may also include effects of a buckling of the crust due to the Caledonian orogeny and emplacement of a thrust sheet to the north and northwest, as suggested by Baarli (1990).

3.1.2. Foreland basin development during the Early Silurian

With the ongoing closure of the Iapetus Ocean and formation of the Caledonides, the regional setting of the Oslo Region changed successively. The shallow epicontinental sea turned slowly into a foreland basin, in front of the Caledonian thrust sheet, highly influenced by increased subsidence due to nappe-loading towards the Middle Silurian (Bjørlykke, 1983) (Fig. 3.3). The sand layers in between the limestone layers got thicker
which can be seen as an index for an increased sedimentation rate in the uppermost Ordovician succession in the Oslo Region connected to the Caledonian orogeny (Bjørlykke, 1974a).

The first response to the development of the Caledonian orogenic belt can already be seen in the Middle Ordovician, fine-grained sandstones of the Elnes Formation (Bruton et al., 2010). More prominent sandstone formations occur during the Late Ordovician, representing clastic influx from erosion of the thrust sheets approaching from the west into the epicontinental basin to the east (Bruton et al., 2010). These sandstone formations show a lot of regional facies changes due to unstable tectonic conditions due to the Caledonian orogeny in the west (Worsley and Nakrem, 2008).

Approximately one half of the ca. 2000 m thick Lower Palaeozoic succession in the Oslo Region (Bockelie and Nystuen, 1985) was deposited during the Cambrian and Ordovician (ca. 100 Ma), whereas the other half of the succession was deposited during the Silurian time which lasted for about 24 Ma (Baarli, 1985; Worsley and Nakrem, 2008).

Even though the sediment supply was higher than in Ordovician times, sedimentary successions recorded from the Silurian indicate an increased water depth (Worsley and Nakrem, 2008), due to the Caledonian nappe loading in the north of the Oslo Region (Bjørlykke, 1983; Baarli, 1990).

3.2. Late Ordovician - Early Silurian Stratigraphy

The stratigraphy of the Langøyene Formation as well as the Husbergøya Formation and Solvik Formation will be shortly summarised to get a more complete overview of the Middle Ordovician and Lower Silurian sediments in the central part of the Oslo Region.

The Upper Ordovician Husbergøya and Langøyene formations were initially described by Brenchley and Newall (1975) and revised by Owen et al. (1990). The succession in the central Oslo Region continues with the Lower Silurian Solvik Formation, first defined in Worsley et al. (1983).

3.2.1. Husbergøya Formation

The base of the Husbergøya Formation is defined above the nodular limestones in the uppermost part of the Skogerholmen Formation, starting with dark grey shales, interbedded with calcareous beds and thin sandstones of increasing frequency and thickness upwards (Brenchley and Newall, 1975; Owen et al., 1990). The uppermost
part of the Husbergøya Formation is defined by a 2 to 5 m thick, brown weathering, bioturbated sandstone (Brenchley and Newall, 1975; Owen et al., 1990). The Husbergøya Formation is 18.5 m thick at its type locality at Husbergøya. The total thickness throughout the Oslo Region is described as relatively constant, ranging between 17-25 m (Brenchley and Newall, 1975). Brenchley et al. (1979); Brenchley and Cocks (1982) described the depositional environment of this formation to be at a deep shelf in the offshore transition (cf. Fig. 4.1).

This was supported by Kjærsgaard (2014), who proposed a depositional environment with stable platform conditions on the Baltoscandian epicontinental shelf, close to storm weather wave base, in a relatively proximal position.

### 3.2.2. Langøyene Formation

The lower boundary of the Langøyene Formation was defined chronostratigraphically by Brenchley and Newall (1975), in contrast to the previous, biostratigraphically defined boundary by Kiær (1897). The formation starts right above the brown, weathering sandstone of the Husbergøya Formation, with a gradational contact towards the

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**Figure 3.4.** Age and stratigraphic relationship of the uppermost Ordovician and lowermost Silurian succession in the central Oslo Region, following the schemes in Worsley and Nakrem (2008) and Bruton et al. (2010), with stratigraphic ages from Cohen et al. (2013 updated). Formation boundaries are plotted in approximate positions, since absolute dates are not available. The relative sea-level curve is adapted from Nielsen (2004).
underlying brown weathering sandstone (Brenchley and Newall, 1975). The lowermost succession of the Langøyene Formation is characterized by shales with interbedded, discontinuous limestone beds and calcareous sandstone which may have erosive bases or some minor ripple structures (Brenchley and Newall, 1975) (Fig. 3.5). There also occur layers with contorted bedding, containing well-rounded, sub-spherical quartz grains a few metres above the lower boundary (Spjeldnæs, 1957; Brenchley and Newall, 1975).

The formation continues with thin-bedded and more massive sandstone beds, containing ‘millet seed’ quartz grains and limestone conglomerate beds (Brenchley and Newall, 1975). Brenchley et al. (1979) and Brenchley and Newall (1980) discussed a storm surge origin of the sand beds during a glacio-eustatic, regressive period in the epicontinental sea on Baltica.

There are up to three conglomerate units which can be found interbedded with the sandstone beds (Brenchley and Newall, 1975, 1980). Their erosive lower boundaries were interpreted to represent bases of channel entrenchments within an incised valley system in a shallow epicontinental sea, with the lower conglomerate beds as their first fluvial infill (Sandbakken, 2014). Other possible origins of the conglomerate units on the erosional surfaces were proposed by Brenchley and Newall (1975, 1980); Brenchley
et al. (1979); Worsley and Nakrem (2008) to be tidal channels, by Baarli (1990) to have formed due to sea-level changes connected to a peripheral foreland bulge in front of the Caledonian thrust belt, and by Braithwaite and Heath (1992), who investigated the Late Ordovician in the Hadeland area (ca. 50 km north of Oslo), as channelized beds of conglomerate deposited as debris flows.

The uppermost part of the Langøyene Formation is often marked by an oolitic, arenaceous, cross bedded limestone (Brenchley and Newall, 1980) (Fig. 3.6). This type of arenaceous limestone was also found as clasts in some of the conglomeratic units in the central Oslo Region, implying that the carbonate unit must have been lithified soon after the deposition (Brenchley and Newall, 1980).

The Langøyene Formation is 51 m thick at its type locality at Langøyene (Brenchley and Newall, 1975), but shows a high grade of lateral variations (Brenchley and Newall, 1980).

The southern sections at Langøyene are dominated by sandstones in which soft sediment deformation structures occur frequently. The total thickness of the formation in this area is 53 m (Brenchley and Newall, 1975).

The northern sections, at Gressholmen, Bleikøya and Hovedøya, show more interbedding of shales and sandstones, much less of deformation structures, and limestone conglomerate beds are often lacking (Brenchley and Newall, 1975).

The Langøyene Formation is relatively thin in a central belt in Asker, with thicknesses between 17 and 25 m, showing a succession of laminated quartz sandstones overlain by oolitic limestones. The northernmost sections are only represented by brecciated oolitic limestone, with thicknesses below 2 m (Brenchley and Newall, 1975). The depositional environment of this formation was characterized by offshore transition to shoreface conditions (Brenchley and Cocks, 1982) (cf. Fig. 4.1).

### 3.2.3. Solvik Formation

The Solvik Formation was defined and first described in Worsley et al. (1983). It is mainly divided in two Members: Myren and Padda Member.

The flooding surface at the top of the underlying brown weathering sandstone from the Langøyene Formation (Fig. 3.6), overlain by dark grey, silty shales or an about 60 cm thick nodular limestone unit defines the lower boundary of the Myren Member and likewise of the Solvik Formation (Worsley et al., 1983).

The thickness of the Myren Member is at least 160 m (Worsley et al., 1983). This lower member of the Solvik Formation is characterized by shale deposits interbedded with very thin siltstones, 1 to 3 cm thick (Worsley et al., 1983).
Figure 3.6: Top of the Langøyene Formation on Hovedøya, exposed at the southern locality on Hovedøya, where the brown weathering sandstone lies directly on top of a subaerial unconformity. Note that the bedding is overturned – the arrow in the lower right corner indicates the stratigraphic right way up.

The Padda Member is lithologically dominated by shales, but contains interbedded lenses and beds of calcareous siltstones and limestones with calcareous nodules (Worsley et al., 1983). The base of this member is defined to be at the first occurrence of these interbeds at a sharp boundary (Worsley et al., 1983). The transition of the Padda Member of the Solvik Formation into the overlying Rytteråker Formation is gradual (Worsley et al., 1983). However, the lower boundary of the Rytteråker Formation is set at the first occurrence of a thin, between 3 and 10 cm thick nodular limestone horizon (Worsley et al., 1983).

Deposition took place under quiet conditions after the Early Silurian transgression, with mainly mud-sedimentation and occasional deposition of storm-generated siltstone layers (Worsley et al., 1983). Throughout the whole Solvik Formation, an increase in energy level can be observed, marked by the appearance of erosive bases in
the Myren Member and storm-lag deposits at the top of the Padda Member (Baarli, 1985). The limestones of the Padda Member were interpreted to indicate shallower conditions than the underlying Myren Member (Worsley et al., 1983).

The depositional environment of the Solvik Formation was described by Baarli (1985) to be in an offshore transition to offshore position (cf. Fig. 4.1).
4 Deposition and Basin dynamics

4.1. Epicontinental basins and shallow seas

Epicontinental or epeiric seas cover continents with water depths between a few tens of meters in coastal areas to a few hundreds of meters in deeper areas and are bordered by land (Einsele, 1992; Nichols, 2009). They have their highest extent during eustatic sea level high stand and are very sensitive to sea-level fluctuations, which may relocate the shorelines significantly (Wellner and Bartek, 2003; Nichols, 2009; Midtkandal and Nystuen, 2009; Glørstad-Clark et al., 2010, 2011). Recent examples for epicontinental seas are the Barents Sea, the Arafura Sea, Hudson Bay, Gulf of Carpentaria (Midtkandal and Nystuen, 2009), the Baltic Sea (e.g. Allison and Wells, 2006) and the Yellow and East China Sea (e.g. Alexander et al., 1991; Uehara and Saito, 2003). The broad and shallow, low-ramp shelves, usually with a dip less than 1° cause very gradual transitions between different facies, which tend to be very extensive over large areas (van Wagoner et al., 1990).

Epicontinental seas are one of two types of shallow seas described in literature; another type is the marginal or pericontinental sea which covers a “normal” shelf environment (Einsele, 1992). The main factors controlling sedimentation in shallow seas are:

- \( a \) amount of sedimentary input into the basin, depending on climate, surrounding rock types, drainage patterns like rivers and associated deltas;
- \( b \) variations in biogenic production (reefs and carbonate production);
- \( c \) energy regime of the specific environment, depending amongst others on water depth;
- \( d \) sea-level fluctuations and
- \( e \) reworking of earlier deposited sediments (Einsele, 1992).

The main depositional processes in epicontinental seas are thought to be of storm or tidal origin, whereas storms are more dominant in regions where the tidal ranges are quite small (Nichols, 2009). The epicontinental seas in ancient times were considerably larger than the ones we can observe nowadays (Allison and Wells, 2006). Allison and Wells (2006) suggested, that sediments at depths of 40 m can be reworked by annual storms with wavelengths of 80 m. The fair weather wave base will be located at much shallower depths.

There are two main types of epeiric seas: the epeiric platform (Tucker and Wright, 1990) and the epeiric ramp (Wright and Burchette, 1998). The former is dominated by carbonate production with little continental influence on sedimentation (Tucker and
Wright, 1990; Allison and Wright, 2005). In contrast, the latter land-attached deposystems on low-gradient ramps may extend over hundreds of kilometres, where tidal influence is dampened by friction and the tidal effect on sedimentation will therefore diminish (Irwin, 1965; Bjørlykke, 1974b; Allison and Wright, 2005). Epeiric ramp settings have broad facies belts with gradual transitions in between them (Wright and Burchette, 1998). The sea floor lies within the zone of storm wave base and should be divided into a more proximal and distal part, dependent on the influences by fair weather waves (Lukasik et al., 2000). Stratified water conditions were proposed for the epeiric ramp setting and this may in turn constrain the carbonate production (Allison and Wright, 2005). A contrasting view is given by Uehara and Saito (2003), who showed from studies in the Yellow Sea, that tidal forces play a significant role in reworking sediments. The development of the tidal-current regime is, however, dependent on the shape of the basin and isolation from the open ocean (Uehara and Saito, 2003; Allison and Wright, 2005).

The accumulation of sediment in an epicontinental basin is relatively high, since sediments cannot be transported into deeper water (Einsele, 1992). Due to the shallow depth of this depositional environment, it can be classified equivalent to continental shelves into shoreface, offshore transition and offshore environments (Fig. 4.1) (Nichols, 2009).

![Figure 4.1: Schematic illustration of the shelf; representing low-gradient slope environments with water depths down to 200 m; the shelf is divided into foreshore, shoreface, offshore transition and offshore, bounded by water energy ranges: mean high and low water as upper and lower boundary, respectively for the foreshore, fair weather wave base (FWWB) as lower boundary of the shoreface, storm wave base (SWB) as lower boundary for the offshore transition and offshore below the SWB (Nichols, 2009).](image)

**Storm influenced sedimentation**

Storms can be recorded in sediments that were deposited between fair weather wave base and storm wave base, since sediments deposited during storms above FWWB are likely to be reworked by other marine processes (cf. Fig. 4.1).
Tempestites (storm-generated deposits) are highly variable in their appearance in respect to the different storm-generated beds and dependent on the shape and type of the basin (Myrow and Southard, 1996). Typical storm deposits reveal the antiformal hummocky and synformal swaley cross-stratification (e.g. Dott and Bourgeois, 1982; Walker et al., 1983). Experiments by Dumas and Arnott (2006) showed that the formation of hummocky and swaley cross-stratification are closely related to each other and happens between fair and storm weather wave base. In accordance with their investigations, hummocky cross-stratification may form in water depths between 13 and 50 m (Dumas and Arnott, 2006). Following this study, swaley cross-stratification may be seen as truncated HCS and will occur above them (Dumas and Arnott, 2006).

### 4.2. Foreland basins

The general definition of a foreland basin is an elongated trough, formed between an orogenic belt and a stable craton due to lithospheric flexure (Watts, 1992; DeCelles and Giles, 1996). Due to flexural subsidence a basin will form, named as foreland basin or foredeep (DeCelles and Giles, 1996), which is bounded on the cratonic side by a peripheral forebulge (Pigram et al., 1989).

Posamentier and Allen (1993) presented a sequence stratigraphic model in which foreland basins can be divided into zone A and zone B, according to their proximality (Fig. 4.2B). Zone A is the area where the rate of subsidence exceeds the rate of eustatic fall, which will be the landward side of the basin margin (Posamentier and Allen, 1993). Zone B is defined as the area where the eustatic fall periodically exceeds the rate of subsidence (Posamentier and Allen, 1993). Consequently will zone B lie seaward of zone A in ramp-type foreland basins (Posamentier and Allen, 1993).

DeCelles and Giles (1996) defined four depozones in foreland basins: wedge-top, foredeep, forebulge and back-bulge (Fig. 4.2A, C), whereas forebulge and back-bulge can be absent in some settings. The main sedimentary accommodation takes place in the foredeep depozone as a result of flexural subsidence due to tectonic or sedimentary load (DeCelles and Giles, 1996). Thereby the thickness of the accumulated sediments decreases towards the forebulge (Horton and DeCelles, 1997).

Accommodation in the back-bulge and forebulge happens due to aggradation up to the equilibrium profile (DeCelles and Giles, 1996). Successions from these depozones are relatively thin compared to the ones deposited in the foredeep depozone (DeCelles and Giles, 1996). Sediments in the back-bulge depozone taper onto the forebulge, as well as the craton (Horton and DeCelles, 1997). The formation of intrabasinal arches has be described and linked to a forebulge zone in an example of a carbonate domi-
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Figure 4.2.: Schematic representation of foreland basins by DeCelles and Giles (1996): A) Schematic map view of a foreland basin, the vertical line represents a cross section that may look like B; B) general accepted cross section of a foreland basin with indication of zones A and B according to Posamentier and Allen (1993); C) schematic cross-section, following the model by DeCelles and Giles (1996) with the four main depozones: wedge-top, foredeep, forebulge and back-bulge.

The sediments derived from the thrust belt, as well as from the forebulge and intrabasinal carbonate sediments, may be deposited over large areas, far away from the major flexural subsidence (DeCelles and Giles, 1996). The length of the foreland basin can be assumed to have approximately an equal length to the adjacent fold-thrust belt (DeCelles and Giles, 1996).

The model described by DeCelles and Giles (1996) (Fig. 4.2A, C) implies that foreland basins are much more various than described in the earlier, wedge-geometry models. Their developed depositional facies vary strongly, depending on the interplay of subsidence and sedimentation in the considered area (Catuneanu, 2008). The thrusting stage and the flexural response of the lithosphere have a strong influence on the geometry of the basin (Watts, 1992). As a response to the thrusting, the foreland basin, as well as the forebulge, migrates towards the craton (Pigram et al., 1989).

Foreland basins occur usually in two different shapes; either wide and shallow or...
narrow and deep (Watts, 1992). The initial fill occurs on a weak crust and will cause therefore a short wavelength flexure resulting in a deep basin (Watts, 1992). As thrusting continues, load moves onto the stronger lithosphere of the craton and the foreland basin will become rather broad and shallow (Watts, 1992).

4.3. Incised Valleys and Estuaries

4.3.1. Incised Valleys

Incised valleys were defined by Zaitlin et al. (1994) as fluvially eroded, elongated topographic lows, which are larger and deeper than one single fluvial channel. Their base is defined to be formed by a regionally mappable unconformity and a basinward shift in facies along that unconformity (van Wagoner et al., 1990; Zaitlin et al., 1994).

Incised valleys are stratigraphically isolated environments, in which allochthonous and autochthonous processes can be observed regarding their influence on sedimentary processes, and therefore are of high importance in the stratigraphic record (Dalrymple, 2006). If sea level falls under a significant topographic break, incision will occur from the downstream end and propagate landward (Catuneanu, 2008). If a low-angle ramp setting is exposed during such a sea-level lowstand, the whole shelf is likely to be incised (Catuneanu, 2008).

There are two main situations that may force the formation of incised valleys: a) the slope of the river increases – either due to a forced regression or due to a differential uplift of the region (conformable with a relative fall in base level) or b) the increase of the ratio of water to sediment discharge, which may happen due to climatic changes, tectonic subsidence or uplift (Dalrymple, 2006). Even though the first of these two main situations is often expected to be the most common type, changes in climate and accompanied changes in vegetation may have an equal influence on depositional systems (Zaitlin et al., 1994; Dalrymple, 2006). Both of the described cases result in an abrupt increase in fluvial energy, which will cause the incision of the river (Catuneanu, 2008). Unconformities in continental settings are often controlled by climatic and tectonic conditions in the source area (Shanley and McCabe, 1994). An autocyclic origin of incised valleys due to switching of distributary channels is discussed by Ainsworth and Walker (1994). The relative importance of all of the mentioned factors varies with both time and space (Shanley and McCabe, 1994). The increase of the slope and the accompanied change in the equilibrium profile is, however, of high sequence stratigraphic significance since regionally extensive unconformities are formed in that way.
and may be interpreted and used as sequence boundaries (Zaitlin et al., 1994).

There are two major types of incised valley systems: the **piedmont incised valley system** (Fig. 4.4, 4.3), which draws its water from a (mountainous) hinterland and crosses a fall line, where there is a significant reduction in slope gradient and the **coastal-plain incised valley system**, which is confined to low-gradient coastal plains (Fig. 4.3) (Zaitlin et al., 1994; Boyd et al., 2006). Piedmont incised valley are often linked to tectonic processes in the hinterland and may therefore be longer-lived than coastal-plain incised valley systems that are associated with relative sea-level changes (Boyd et al., 2006). The outcome of this fact is that piedmont incised valley systems are characterized by a longer fluvial reach than coastal-plain incised valley systems, and their catchment areas are therefore usually larger with a higher sediment supply (Boyd et al., 2006). The initial infill of the former type will often consist of immature fluvial coarse-grained sediment whereas the last one rather consists of finer-grained and more mature sediments, reworked from the coastal plain (Zaitlin et al., 1994; Boyd et al., 2006).

In coastal systems, the presence or absence of rivers is a crucial factor for the development of sedimentary facies (Boyd et al., 1992). Rivers will incise into existing topography and follow pre-existing lows (Dalrymple, 2006). Compound fills may occur, where older incised valleys still have unfilled accommodation space (Zaitlin et al., 1994). Fluvial systems produce a network through the incised valley that will either contribute to the erosion and formation of the incised valley, or supply sediment for the very first valley infill during initial relative sea-level rise (Zaitlin et al., 1994).

Zaitlin et al. (1994) proposed a tripartite segmentation of incised valleys (Fig. 4.6):

![Diagram of incised valleys](image-url)
Figure 4.4.: Development of a simplified piedmont incised valley system through a complete sea-level cycle: A) Lowstand systems tract (LST) with the formation of the incised valley system; the point where the incised valley system is passing into the non-incised river system is called knickpoint; B) LST with a delta at the mouth of the incised valley, beginning fluvial deposition inside the incised valley; C) Transgressive systems tract (TST), formation of the tripartite zonation inside the incised valley system, in this figure; D) Highstand systems tract (HST) with a prograding shoreface and a coastal plain where the incised valley was buried (Zaitlin et al., 1994).
Segment 1, as the outer part of the incised valley, characterized by backstepping fluvial and estuarine deposits (lowstand to transgressive succession), overlain by marine sands and shelf muds; Segment 2, as the middle part of the incised valley, showing fluvial to estuarine sediments (lowstand to transgressive succession), followed by a drowned estuarine succession; Segment 3 in the inner part of the estuary, consisting of fluvial deposits.

The cross section of incised valleys is, amongst others, dependent on the duration of how long the river occupied the valley and how much sediment load it carried (Dalrymple, 2006). Complex cross sections are preferably formed by rivers that occupied the valley throughout the falling stage and more simple cross sections can be expected if the rivers occupied the valleys for a relatively short time and had a large sediment load (Dalrymple, 2006). Nevertheless, the valley shape is dependent on the topographic relief as a result of different resistance to erosion (Dalrymple, 2006). The geometry of the sequence boundary at the base of the incised valley is of high importance for palaeogeographic reconstructions amongst others in terms of palaeodrainage directions (Zaitlin et al., 1994).

The facies abundance of the incised valley infill is mainly dependent on the following factors: a) the relative position to the trajectory of the shoreline, b) the ratio of accommodation space to sediment supply, which is dependent on eustatic sea level changes, subsidence- or uplift-rates, the size of the drainage basin and the climate, and c) the rate of shoreline transgression, depending on the ratio of accommodation space to sediment supply and the slope of the surface that is transgressed (Dalrymple, 2006). The trajectory of the shoreline was defined by Helland-Hansen and Martinsen (1996) to represent the cross-sectional path of the shoreline as the basin fills. Marine sedimentation processes become more important if there is little sediment supply from rivers (Dalrymple, 2006).

4.3.2. Estuaries

There are two co-existing definitions of estuaries: the geologic definition, based on the physical processes operating in fluvial-marine transition, producing facies influenced by tides, waves and fluvial processes, and the assumption of a landward migration of facies (Dalrymple et al., 1992) and the oceanographic, salinity-based definition, describing an estuary as a semi-enclosed coastal embayment with brackish water, caused by dilution of seawater with fresh water (Pritchard, 1967).

An estuary forms during transgressions when the rate of relative sea-level rise ex-
ceeds the level of sediment supply, embayed in palaeovalleys (e.g. incised valleys) (Boyd et al., 1992). In this work, the term estuary will be used according to the geologic definition and only their depositional dynamics will be shortly reflected in this chapter.

The geologic definition of estuaries by Dalrymple et al. (1992) implies the existence of an incised valley with a sequence boundary at its base, followed by a transgressive stacking pattern (Dalrymple, 2006). This definition may cause trouble due to the required (incised) valley and the sequence boundary at its base to fulfill it, whereas these conditions are according to Dalrymple (2006) not necessary for the definition of an estuary. More important is the transitional position between land and sea where terrestrial and marine environments interact and the landward sediment transport driven by tidal asymmetry (Cooper, 2001; Dalrymple, 2006).

All these arguments lead to the improved definition, in which an estuary is described as a transgressive coastal environment at the mouth of a river with sediment deriving from fluvial and marine sources and sedimentary facies controlled by tidal, wave and fluvial processes (Dalrymple, 2006).

Figure 4.5.: Tripartition of wave- and tidal dominated estuaries. Figure modified from Nichols (2009) with adapted model of tripartite estuaries by Dalrymple et al. (1992).
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Estuaries show a tripartition in their structure: the outer part, influenced by marine processes (waves and tides); the central part, characterized by the interaction of fluvial and marine processes and the inner part, characterized by fluvial processes (Fig. 4.5) (Dalrymple et al., 1992). Simplified one can say that they extend from the landward limit of the tidal facies to the seaward limit of coastal facies (Fig. 4.5) (Dalrymple, 2006). Estuaries can be subdivided in wave- and tide-dominated estuaries (Fig. 4.5), as well as non-barred estuaries that are characterized by higher energy conditions within the estuary due to the lack of barriers (Cooper, 2001). The former two types are characterized by lower energy conditions behind their barriers (Cooper, 2001).

In an idealized tide-dominated estuary, most of the sedimentary infill comes from tidal currents (Hori et al., 2001), which also contribute to reworking of sediments in the central parts of the estuary (Dalrymple et al., 1992). There are five main facies occurring in an tidal-dominated estuary infill succession: the tidal river, distributary channels, muddy intertidal to subtidal flats, transgressive lags and estuary fronts (Hori et al., 2001). From this perspective one can expect a general fining upwards trend in the succession (Hori et al., 2001).

Seasonal climate may also have a strong influence on the development of the estuary in terms of transport capacity of fluvial or wave energy from marine processes or erosivity, vegetation and slope gradient in the hinterland (Cooper, 2001).

4.3.3. Important stratigraphic surfaces in the estuarine infill of incised valleys

The most important stratigraphic surface in an incised valley system is the subaerial unconformity, also termed as sequence boundary (Fig. 4.6C) (Zaitlin et al., 1994). It is formed through an interplay of fluvial erosion and subaerial exposure of the interfluves during a relative sea-level lowstand (Fig. 4.4A) (Zaitlin et al., 1994). The sediments overlying the sequence boundary may be assigned to a lowstand systems tract (LST) (Fig. 4.3B, 4.6B) (Zaitlin et al., 1994; Dalrymple, 2006) but in cases where sediments were reworked these sediments are rather assigned to a transgressive systems tract (TST) (van Wagoner et al., 1990; Ainsworth and Walker, 1994). The LST may also include fluvial deposits since the accumulation of those sediments is unlikely during a fast transgression (Allen and Posamentier, 1993).

The TST onlaps onto the LST and this onlapping surface is described as the transgressive surface (Fig. 4.6C) (Allen and Posamentier, 1993). The transgressive surface usually occurs low in the infill succession and forms the main part of the incised valley infill (Allen and Posamentier, 1994; Zaitlin et al., 1994). The TST (Fig. 4.4C, 4.6B) is charac-
terized by changes from fluvial to coastal to marine facies environments (Fig. 4.6A),
developed as retrogradational parasequence sets (van Wagoner et al., 1990), clearly
showing a landward shift of facies (Dalrymple, 2006). Sedimentary facies in the TST
typically represent estuarine sands and muds (Allen and Posamentier, 1993). The first
contact of fluvial to estuarine sediments in the outer part of the incised valley repre-
sents the initial flooding surface that is likewise a transgressive surface (Zaitlin et al.,
1994). The boundaries of backstepping parasequences can be described as flooding sur-
faces, which indicate the intermittent character of the transgression (Zaitlin et al., 1994),
but may also be connected to changes in sediment supply (Shanley and McCabe, 1994).

There may occur a diachronous tidal ravinement surface (Fig. 4.6C), which typically
occurs within the TST (Allen and Posamentier, 1993; Zaitlin et al., 1994). If there is a sig-
nificant tidal influence, this surface will appear with locally eroded channels (Zaitlin
et al., 1994) cutting into the underlying initial estuarine sediments (Allen and Posa-
mentier, 1994). It may not be correlated in large areas, which helps to distinguish
them from fluvial incised channels (Zaitlin et al., 1994). The tidal ravinement surface
may, however, be regionally extensive in settings with a high tidal influence (Allen and
Posamentier, 1993).

Unlike the tidal ravinement surface, the wave ravinement surface (Fig. 4.6C) can be
traced regionally and its morphology is relatively planar (Zaitlin et al., 1994). Above
this surface one can expect transgressive successions that occur in retrogradationally
stacked parasequences (landward migration) (Allen and Posamentier, 1994; Zaitlin
et al., 1994), mainly consisting of estuarine sands (Allen and Posamentier, 1993).

The maximum flooding surface (MFS) represents the time of maximum transgression
throughout the incised valley and divides the TST from the regressive systems tract
(RST) (Fig. 4.6C) (Zaitlin et al., 1994). This boundary may be found within estuarine
deposits, a bayhead delta in the central part of the incised valley system, representing
progradation during initial regression and within fluvial succession, it will be repre-
sented by the sediments with the most distal character (Zaitlin et al., 1994).

Highstand systems tracts (HST) (Fig. 4.4D, 4.6B) are in most settings less well de-
veloped, since too little sediment supply may lead to the direct transition from TST
to RST (Allen and Posamentier, 1993). The HST appears typically as an aggradational
parasequence set in its early stage and changes to progradational parasequence sets
(van Wagoner et al., 1990).
Figure 4.6.: Simplified and idealized longitudinal section of an incised-valley system: A) depositional environments; B) systems tracts and C) important stratigraphic surfaces (Zaitlin et al., 1994).
5 Methodology and Material

5.1. Field work

The exposures of the Langøyene Formation at Hovedøya, Rambergøya and southern Langøyene have been the central working area for this studies (Fig. 3.1).

Detailed sedimentary logs were drawn in scales 1:50 at the western side of Rambergøya and Hovedøya and in scale 1:100 at the eastern side of Rambergøya, at the southern island of Langøyene both in southwestern and northeastern end. The grain size of the sandstones in the logged sections was determined optically in the field by using grain size charts. The logs from Rambergøya cover stratigraphic sections of 40 and 72 m at the eastern and western side, respectively. The log from Langøyene west covers a stratigraphic interval of 70 m, and from the east 61 m. At Hovedøya there were eight logs drawn at the southeastern side of the island, parallel shifted towards east to get a detailed reconstruction of one incised valley structure (Fig. 5.1). Additionally there was a short section logged approximately in the middle of the southern coast of Hovedøya directly under the subaerial unconformity (cf. Fig. 3.1).

Main parts of the sedimentary logs, as well as details within the layers were documented with photographies, using a “Panasonic Lumix, DMC-TZ25”.

The thickness of the Langøyene Formation in the different profiles is shown in Table 5.1.

There were rock samples taken on all of the considered islands either with hammer and chisel, or by collecting loose rocks from respective layers, where work was done.

<table>
<thead>
<tr>
<th>Locality</th>
<th>thickness [m] of the Langøyene Formation</th>
</tr>
</thead>
<tbody>
<tr>
<td>Rw</td>
<td>52</td>
</tr>
<tr>
<td>Rø</td>
<td>35</td>
</tr>
<tr>
<td>Lw</td>
<td>63</td>
</tr>
<tr>
<td>Le</td>
<td>61</td>
</tr>
<tr>
<td>H</td>
<td>46</td>
</tr>
</tbody>
</table>
in protected areas as at Hovedøya. Permission to take samples from protected areas at Gressholmen and Rambergøya was given by the County Governor of Oslo and Akershus.

The logged section on the western side of Rambergøya is the reference section for this work, since it is the most accessible of the complete stratigraphic sections of the Langøyene Formation in the considered area.

Rock samples on Rambergøya were taken every second meter throughout the Husbergøya and Langøyene Formation for isotopic analysis. Samples from Langøyene were taken with irregular spacing to be correlated with the samples from the reference section. Samples from Hovedøya were collected around the conglomerate beds in the southeastern part of the island and below the subaerial unconformity at the central southern shore.

In total there were 38 samples taken on Rambergøya (Tab. G.1), one on Gressholmen, 17 on Langøyene (Tab. G.2) and 6 collected on Hovedøya.

**5.2. Laboratory work**

**5.2.1. Rock sample preparation**

All rock samples were cut perpendicularly to the bedding surfaces with a water-based diamond-blade saw and polished with abrasive paper of 120, 220, 300 and 500 grit. After polishing the surfaces were scanned with a "Canon CanoScan 9000F Mark II".
5.2.2. Stable isotope analysis

Samples for stable isotope analysis ($\delta^{13}$C and $\delta^{18}$O) were drilled from the polished slabs, with preference on micritic, fine-grained carbonate.

There were two sorts of micro-drills used for producing the powder for the analysis: a BOSCH HSS-R with 118° lead angle and a diameter of 2 mm and a HSS-G 135° lead angle and a diameter of 1,5 mm. The former (BOSCH HSS-R) gave better results on harder rocks.

Stable isotopes are usually measured from biogenic carbonate structures, such as brachiopods. The availability of fossils throughout the profile was very low in this study and measurements were therefore taken from the micritic matrix of the sand- or siltstones. Spots to drill out powder were determined where there were little amounts of other grains and mainly micritic carbonate matrix.

76 carbonate samples were analysed for stable isotopes ($\delta^{13}$C and $\delta^{18}$O) at the Institute for Geosciences, University of Bergen. Powdered samples were analyzed on Finnigan MAT 251 and MAT 253 mass spectrometers coupled to automated Kiel devices. The data are reported on the VPDB scale calibrated with NBS-19.

5.2.3. X-ray fluorescence (XRF)

Data acquisition

All samples from Rambergøya have been used for a XRF analysis throughout the Langøyene Formation. The measurements were recorded with a HHXRF analyzer ”Therme Scientific Niton XL3t GOLDD+” on cut and polished surfaces with an aperture of 8 mm. Integration time was 90 s in the “Testall Geo” mode.

Chemical background and used proxies

Zirconium (Zr) and silicium (Si) are usually linked to coarser silt and sand fractions in the siliciclastic sediments (Kylander et al., 2011).

Kylander et al. (2011) showed in their study that changes in Zr/Rb can be used as a proxy for changes in grain size, because rubidium (Rb) is more associated to clays, very fine and fine silts, whereas Zr is associated with very fine and fine sands. The ratio Zr/Rb is expected to be lower in fine-grained sediments and relatively higher in coarse-grained sediments (Dypvik and Harris, 2001), as long as the K-feldspar content of the rocks is low because they may contribute significantly Rb (Kylander et al., 2011).

The (Zr+Rb)/Sr ratio indicates the balance between carbonate and siliciclastic components, whereas high values indicate little carbonate contents (Dypvik and Harris,
Titanium (Ti), occurring often in combination with Rb, is also associated with finer grain size fractions (Kylander et al., 2011) and therefore expected to decrease with increasing grain size. It is important to compare only sediments with the same provenance area since the chemical abundance of the measured elements is highly influenced by the chemical composition of the rocks and will therefore vary in different source areas (Dypvik and Harris, 2001). The content of strontium (Sr) is closely associated to the occurrence of calcite and aragonite in the sedimentary succession (e.g. Taylor, 1965; Hammer et al., 1990), due to the often occurring substitution of Sr for Ca in the carbonate lattice (Dypvik and Harris, 2001). This means that Ca and Sr can be used as a proxy for carbonate weathering in catchment areas and likewise carbonate production, in particular CaCO$_3$ and SrCO$_3$ (Kylander et al., 2011).
6 Descriptive sedimentology and depositional environments

6.1. Facies

Facies A

*Description* The facies is characterized by thin shale layers, occurring in beds and laminae of 30 down to a few centimetres in total thickness. The average grain size of rocks in this facies is clay to silt, whereas a minor content of very fine sand may occur. The beds appear to be intensely bioturbated: horizontal burrows are predominant. In some few places bryozoans occur (Fig. 6.1). The shale deposits belonging to this facies are of black to grey colour, appearing brownish after weathering (Fig. 6.12).

*Interpretation* The fine grain size of this facies indicates a depositional environment characterized by low energy conditions. The occurrence of horizontal bioturbation structures shows that the depositional rate must have been relatively low since the considerable reworking of sediments by organisms presumes a long time of non-deposition.

![Figure 6.1](image)

Figure 6.1.: Bryozoa indet. in facies A, 1 m above the base of the Langøyene Formation at the western locality on Langøyene.
### Table 6.1: Overview of facies occurring in the Langøyene Formation

<table>
<thead>
<tr>
<th>Facies</th>
<th>Lithology</th>
<th>Short description, structures</th>
<th>Depositional conditions</th>
</tr>
</thead>
<tbody>
<tr>
<td>A</td>
<td>shale</td>
<td>bioturbation, body fossils, black to grey</td>
<td>low energy and sedimentation rate</td>
</tr>
<tr>
<td>B</td>
<td>sandstone</td>
<td>5 to 10 cm thick beds, ripple or parallel lamination, structureless, dewatering structures</td>
<td>middle energy conditions, rapid deposition</td>
</tr>
<tr>
<td>C</td>
<td>sandstone</td>
<td>20 to 40 cm thick beds, HCS, SCS, planar x-strat, current ripples, low-angle x-strat, climbing ripple lamination</td>
<td>high energy, storm</td>
</tr>
<tr>
<td>D</td>
<td>sandstone</td>
<td>planar x-strat, increased sediment and energy supply</td>
<td>bottom outlet near FWWB, moderate sediment supply and sea level, storm</td>
</tr>
<tr>
<td>E</td>
<td>sandstone</td>
<td>incised channel, low-angle x-strat, climbing ripple lamination</td>
<td>increased sediment and energy supply, storm</td>
</tr>
<tr>
<td>F</td>
<td>sandstone</td>
<td>current ripples, low-angle x-strat, climbing ripple lamination</td>
<td>high energy, storm</td>
</tr>
<tr>
<td>G</td>
<td>sandstone</td>
<td>incised channel, low-angle x-strat, climbing ripple lamination</td>
<td>high energy, storm</td>
</tr>
<tr>
<td>H</td>
<td>sandstone</td>
<td>bioturbation, brown weathering color</td>
<td>moderate sediment supply, bottom currents, above FWWB</td>
</tr>
<tr>
<td>I</td>
<td>conglomerate</td>
<td>clast supported, cut-and-fill structures</td>
<td>high energy, fluvial</td>
</tr>
<tr>
<td>J</td>
<td>sandstone</td>
<td>lensoid-shaped coarse sandstone bodies</td>
<td>high energy, storm</td>
</tr>
<tr>
<td>K</td>
<td>sandstone</td>
<td>trough x-bed, qb-grains, ooides, clasts</td>
<td>middle to high energy, fluvial</td>
</tr>
<tr>
<td>L</td>
<td>sandstone</td>
<td>matrix-supported, coarse sandstone, well-rounded, clasts at lower boundaries</td>
<td>middle energy, fluvial</td>
</tr>
<tr>
<td>M</td>
<td>conglomerate</td>
<td>boulder-conglomerate</td>
<td>undercut carbonate platform deposits</td>
</tr>
<tr>
<td>N</td>
<td>sandstone</td>
<td>coarse sandstone, occasional lithoclasts or alumina</td>
<td>high energy, storm</td>
</tr>
<tr>
<td>O</td>
<td>coquinas</td>
<td>abraded bioclasts</td>
<td>high energy, storm</td>
</tr>
<tr>
<td>P</td>
<td>soft sediment deformation structures</td>
<td>brown, weathering color</td>
<td>bottom stress induced by storms, storm</td>
</tr>
</tbody>
</table>

**Legend:**
- HCS - hummocky cross-stratification
- SCS - swaley cross-stratification
- qz - quartz
- FWWB - fair weather wave base.
Facies B

Description This facies consists of 5 to 10 cm thick sandstone beds, which appear in some places structureless, but may also show current ripple lamination at its base, passing upwards into parallel lamination. The predominant grain size is very fine sand. The lower boundary of beds of this facies appears to be erosive in most places. Dewatering structures occur in some localities. The sandstone beds show a slightly brownish or grey weathering colour and are grey or black on fresh surfaces (Fig. 6.12).

Interpretation The sandstone deposits of facies B are thought to have a storm-origin, likely deposited under middle energy conditions, relatively distal to the area of maximum energy during storm events. Dewatering structures indicate a rapid deposition. Rapid deposition is in favour of that the sand beds were formed during storm events of very rapid sediment transport and deposition.

Figure 6.2.: Facies C, 2,5 m above the base of the Langøyene Formation at the locality on western Rambergøya. The red star indicates the position in the sedimentary log.

Facies C

Description Facies C consists of 20 to 40 cm thick sandstone beds. The sandstone units show planar cross-stratification and interdigitate with each other, forming channel-like structures (Fig. 6.2, 6.14C). Lower boundaries of the beds are developed as erosional unconformities. Downlapping beds occur locally onto the erosional surfaces. Within these sandstone beds, carbonate cemented horizons are locally present. No bioturbation or fossil remains have been observed within the beds, only a few horizontal ones on top of some of the beds. The sandstone units of this facies have a brownish weathering colour and are grey to black on fresh surfaces. Carbonate cemented beds appear more grey, both in weathering and fresh surfaces.
**Interpretation** The planar cross-stratification point to a high-energetic origin of the beds representing facies C. The channel-shaped structures have been formed by erosion brought about by turbulent flow. The lack of internal bioturbation may be due to relative high sedimentation rates or high energy conditions.

**Facies D**

**Description** Facies D consists of 10 to 40 cm thick sandstone beds of fine sand grain size. There occurs hummocky cross-stratification, swaley cross-stratification or parallel lamination. Grading is very common in this facies (Fig. 6.3). In some cases these beds are amalgamated, consisting of up to 6 separate sandstone graded beds, forming composite bed sets. Quartz grains and ooides occur in some of the beds. This facies appears in a light to dark grey colour (Fig. 6.12).

**Interpretation** Hummocky cross-stratification and swaley structures indicate a storm-origin of this facies. The occurrence of quartz grains, which are mainly absent in the surrounding facies, indicates that the sediments originated in a depositional environment characterized by a high energy flow regime. Amalgamated beds point to an increased frequency of storms, since fine-grained background sediments are absent between sandstone beds (Dott and Bourgeois, 1982).

**Figure 6.3.** Facies D, showing amalgamated beds with parallel lamination, ca. 16 m above the base of the Langøyene Formation at the western locality on Langøyene. The red star indicates the position in the sedimentary log.
Franziska Franeck

Figure 6.4: Facies E, showing beds with planar cross-bedding, above the first conglomerate unit, 18.5 m above the base of the Langøyene Formation at the western locality on Rambergøya. The red star indicates the position in the sedimentary log.

Facies E

Description This facies consists of fine- or middle-grained sandstone, showing planar cross-stratification. The sand appears to be relatively well sorted, consisting of well-rounded quartz-grains. The beds representing this facies have erosive, lower boundaries, are graded and reach thicknesses up to 20 cm (Fig. 6.4). Clasts of corals (Fig. 6.4) may occur, but are very rare. The beds representing this facies are of grey or ochre colour.

Interpretation Sandstone units of this facies occur close to conglomerate beds (Facies I) and are therefore thought to be related to phases with increased energy and sediment supply. The presence of well sorted and rounded quartz grains and the planar cross bedding point to deposition under middle to high energy conditions as large 2D sand dunes.

Facies F

Description This facies consists of very fine to fine sandstone, occurring in beds varying from 15 to 50 cm in thickness. Within the beds, current and climbing ripple lamination often occur at the base and are overlain by low-angle cross-stratification (Fig. 6.13). All these internal structures may occur together, but not in all places. Beds with current ripple lamination at the base passing into low-angle cross-stratification are most common. The weathering colour of these beds is ochre to greyish and dark grey on fresh surfaces.

Interpretation The ripple structures and the low-angle cross-stratification indicate the deposition under high energy conditions. The climbing ripple lamination points
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Figure 6.5.: Facies G, 25 m above the base of the Langøyene Formation at the locality on western Rambergøya. The red star indicates the position in the sedimentary log.

to high sediment supply Ghienne (2003). The sedimentary structures in beds of this facies indicate that the beds were deposited by unidirectional flows. Unidirectional flow pattern is typical for storm events and rarely occurs under fair weather conditions in shelf or ramp settings (Dumas and Arnott, 2006). The internal structures therefore point to that individual beds of this facies represent separate single depositional events, which may have been a storm events.

Facies G

Description The laminae in this facies are up to 5 cm thick and consist of very fine to fine sandstone (Fig. 6.13). There are small symmetrical wave ripple lamination on the top of the laminae (≤1 cm high, Fig. 6.5). The weathering colour of these rocks is dark grey.

Interpretation The laminae of this facies appear to have been deposited slightly above the fair weather wave base (FWWB), where oscillatory wave-induced bottom currents were able to form ripple lamination. Fine sand laminae may be related to moderate sedimentation rates, which also are supported by the occurrence of bioturbation on the top of some layers.

Facies H

Description This facies comprises massive, structureless sandstone beds, consisting of very fine sand. The thickness of the sandstone bed varies from 1 to 5 m in
the studied localities. The sandstone beds are highly bioturbated and contain a considerable amount of body fossils, as bryozoans, trilobites, brachiopods and cornulites. The weathering colour of this facies is brown to ochre, and grey on a fresh surface.

**Interpretation** Deposits of this facies represent a condensed section, which can be inferred from the complete reworking by bioturbation. The grain size and bioturbation indicate that the water depth likely was below the FWWB, but not very deep. Sandbakken (2014) showed in his study that the brown weathering color of this facies is due to the occurrence of ankerite (CaFe\(\text{CO}_3\)) and dolomite (CaMg\(\text{CO}_3\)) minerals in the sandstone.

**Facies I**

**Description** This facies is defined as a clast-supported conglomerate (Fig. 6.14A). Bed thicknesses vary from 20 to 70 cm, and individual beds extend laterally up to 7 m. Their lower boundaries are erosive and internal grading occurs frequently. The beds appear as cut-and-fill structures. The clasts in the individual beds are well-rounded, elongated and composed of both litho- and bioclasts. Whereas lithoclasts represent the underlying layers, bioclasts are made up of coral fragments that are not very common in the underlying strata. Clasts occurring together in the same depositional unit are of similar grain size class, varying from some centimetres up to 20 cm or longer in their longest axis. This facies may pass upwards into cross or planar bedded sandstone or coarse sand lenses (cf. facies E and J, Fig. 6.14A). The colour of the matrix is light grey to grey and the clasts are mainly darker grey and brownish weathering, in case of bioclasts is the colour light grey.

**Interpretation** This facies is interpreted to have been deposited in fluvial environments. Cut-and-fill structures were formed by fluctuating energy in the current from being erosive to be depositional. Well-rounded clasts indicate a rather long total transport of the clasts.

**Facies J**

**Description** This facies occurs commonly associated with facies I and includes coarse-to middle-grained sandstone bodies (Fig. 6.14A). The beds are lensoid-shaped, up to 40 cm thick and extend laterally up to 5 m. The lower boundaries are erosive. The beds are graded and show in some places cross-stratification. Fossils are not abundant, but may appear occasionally in the coarser fractions and are
there represented by coral fragments. The sandstone beds are of light grey to brownish grey colours.

**Interpretation** This facies has been interpreted to have been formed through fluvial processes. The cut-and-fill structures may represent sand dunes or minor bars in highly mobile braided river systems. Compared to the associated facies I, the smaller grain size in this facies may point to less flow energy, due to a shift in the main-drainage pattern, less availability to coarser-grained debris or rather better reworking and sorting of sediments.

![Figure 6.6: Facies K, exposed at the southern locality on Hovedøya. Note that the beds are overturned and therefore the photograph is turned upside down. The predominant transport direction of sediment has been towards west. The red star indicates the position in the sedimentary log.](image)

**Facies K**

**Description** This facies is defined as trough and planar cross-beded sandstone (Fig. 6.6), consisting of calcareous sandstone with a relatively high amount of quartz grains. The grains are very well rounded and of fine middle to coarse grain-size. Clasts consisting of coral fragments occur frequently at the base of troughs. The sandstone of this facies shows a middle to dark grey colour.

**Interpretation** Trough cross-bedding indicates deposition in a high energy environment. Climbing dunes indicate not only deposition under high energy, but also
a high amount of sediment deposited in a relative short time. The rather small grain size and well-rounded shape of clasts point to middle to high energy depositional conditions.

**Facies L**

**Description** This facies consists of middle- to coarse-grained sandstone beds with clasts at the base of the beds. The conglomeratic base passes into cross-bedded sandstone showing less intense cut-and-fill structures (facies C). Cross-stratification of dunes and climbing dunes may occur in some localities (Fig. 6.14B). The sandstone beds are cross-bedded and show cut-and-fill structures, whereas the single beds are about 30 to 40 cm thick (Fig. 6.14C). The clasts consist of rocks that can be found in the strata below the first major unconformity within the Langøyene Formation. The colour of the sandstone is grey, whereas the clasts mainly appear in brownish weathering colours.

**Interpretation** The conglomerate beds contain clasts that were most likely eroded from underlying strata, followed upwards by sand that was washed in by fluvial processes. The conglomeratic lower part, upper cross-bedded part and the coarse sand grain size of the sandstone beds point to a fluvial environment. The occurrence of dunes and even climbing dunes indicate deposition in a high energy depositional environment.

**Facies M**

**Description** This facies is a conglomerate that consists of huge clasts (m-scale) above an erosional unconformity (Fig. 6.7). The clasts consist of cross-stratified carbonate cemented sandstone (cf. facies K) and may have diameters up to 2 m. The clasts are abraded at the edges and appear in a block-like shape, but there are no signs of significant transport. Smaller clasts are very rare and if they occur, they are included in the fine- to middle-grained sandstone matrix between the huge carbonate sandstone blocks.

**Interpretation** Since the blocks in this facies are huge, it is unlikely that they have been transported very far. The carbonate cemented sandstone of the blocks has most likely been a sand bar of platform sediments, generally dominated by carbonate deposition. By early carbonate cementation of the sand, the sand bar deposit became resistant to erosive forces. Underlying strata and weak areas were eroded, likely by a stream entrenching the platform (discussed below). The carbonate-
cemented platform sandstone was undercut and broke apart in large blocks that fell into the channels.

**Figure 6.7.** Facies M at the western locality on Hovedøya. Large, elongated carbonate-sandstone boulders are laying on top of the erosive lower boundary. Note that the beds are overturned and the stratigraphic right way up points downward. The red star indicates the position in the sedimentary log.

**Facies N**

**Description** This facies consists of middle to coarse sandstone beds, deposited on an erosive lower boundary. Clasts of up to 7 cm in size occur locally at the base of the sandstone beds (Fig. 6.8). The beds are up to 50 cm thick and pinch laterally out. In thicker beds (20 to 50 cm), the clasts occur mainly as lithoclasts, whereas thinner beds contain mainly bioclasts, consisting of coral or crinoid fragments (Fig. 6.8B).

**Interpretation** According to the coarse grain size and clasts, these beds have most likely been deposited under relatively high energy conditions. They may have been formed during especially large storm events. There occurs a change of predominant clasts upwards in the strata from lithoclasts to bioclasts. This may be an effect of available material that was eroded somewhere else.

**Facies O**

**Description** This facies consists of limestone beds consisting predominantly of bioclasts. The beds have an erosive lower boundary and are 5 to 10 cm thick. The
Figure 6.8.: Facies N at the western locality on Hovedøya. A: ca. 37 m above the base of the Langøyene Formation, showing coarse-grained sandstone without any clasts; B: ca. 41 m above the base of the Langøyene Formation with numerous abraded rugose coral fragments within coarse-grained sandstone. The red stars indicate the positions in the sedimentary log.

Main components of the clasts are composed of gastropods, echinoderms, anthozoans and brachiopods (Fig. 6.9). The beds are grey in colour due to the high calcareous content.

**Interpretation** The erosive lower boundary of the beds in this facies point to a clear high-energy origin of this facies. Abraded fossil fragments indicate transport of the clasts.

Coquina beds from other localities have been described from other localities to consist of a high amount of bioclasts that were deposited under high-energy conditions (Brenner and Davies, 1973). The thickness of the beds in this facies is relatively thin and therefore the beds can be expected to be of local extend. This emphasizes the hypothesis of a storm-origin of this facies (Brenner and Davies, 1973) since these events are restricted locally. The bioclasts have most likely been eroded in a more tranquil and protected area during storm events and washed
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together as coquina beds.

Figure 6.9.: Coquina bed; facies O on Hovedøya with some examples of occurring fossils: B brachiopod, C coral, Crin crinoid fragment, Gast gastropod. Detail pictures are from the same bed, close to where the large-scale photographs were taken. The red lines indicate the erosive lower boundary. The red star indicates the position in the sedimentary log.

Facies P

**Description** This facies is defined as soft sediment deformed sandstone beds. The deformation structures are in some places developed as ball-and-pillow structures in meter scale, and in other places as relatively thin contorted beds, tens of centimetres thick (Fig. 6.10, 6.11, 6.12). The grain size varies from coarse sand to fine sand. Deformation structures occur in single-beds that vary in thickness from tens of centimetres to several meters, but may also occur in tens of centimeter-thick bedsets.

**Interpretation** The lithology of deformed beds and bedsets of facies P corresponds to that of adjacent undeformed strata. The deformation structures have formed as a result of collapse of internal coherence and stratal framework of sandstone
Figure 6.10: Soft sediment deformation (convolute lamination) 3 m above the base of the Langøyene Formation on Langøyene as a photography and a schematic sketch of the outcrop. There seems to be several deformation phases or very local occurring early-diagenetic lithification in beds that resist later deformation events. The red star indicates the position in the sedimentary log.

Bottom stress induced by storms may have initiated the deformation of un lithified strata.

Brenchley and Newall (1977) suggested reversed density as a reason of the inversions and contortion of sediment beds. The deformation process may have been triggered by rapid sediment deposition, breaking waves, slope failure due to oversteepening, or deformation due to recurrent stress events, as e.g. earthquakes, storm waves (Owen, 1987). Earthquakes may be a less common trigger for the deformation of sediments than commonly assumed, since they are relatively rare events compared to tidal forces (Greb and Archer, 2007) or seasonal storm events (Alfaro et al., 2002). If deformation structures occur frequently, their triggering agents are likely to be a common event in the specific setting (Molina et al., 1998). The frequent occurrence of sandstone beds with internal structures indicating storm events would support the hypothesis they have triggered soft sediment deformation structures recorded in the Langøyene Formation.
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Figure 6.11.: Soft sediment deformation (convolute bedding, ball-and-pillow structures) 19 m above the base of the Langøyene Formation on Langøyene. The red star indicates the position in the sedimentary log.

6.2. Facies associations

In the following chapter, the terminus “shelf” is used in a palaeobathymetric sense to characterize supposed water and energy conditions. As described before, the depositional environment is a shallow, epicontinental sea on Baltica.

6.2.1. Facies association 1 (FA 1) – outer shelf

Description

This facies association includes the lowermost parts of the Langøyene Formation in all studied localities. It combines facies A, B, C, D and P. The background sediment consists of silt or very fine sand (facies A), in which horizontal bioturbation may occur. The siltstone successions are interrupted by very fine sand- to fine sandstone laminae, which often show an erosive base (facies B). The transition from facies B to facies A is rather sharp, even though grading occurs within facies B beds and laminae, and the generally sharp boundary is unclear in some places. The internal structure of the siltstone beds and laminae is often characterised by parallel lamination and current ripple lamination, even hummocky cross-stratification (facies D) may occur. Sandstone beds with channel-shaped geometry seen in 2D section (facies C, Fig. 6.2) have been observed in most of the studied localities in the lower part of the Langøyene Formation. Soft sediment deformation structures (facies P) have been recorded, often in context with sandstone beds of relatively coarse grain size, compared to the background sediments.
Table 6.2.: Overview of facies associations with interpretations of the depositional environments; FA - facies association; Lithology abbreviation: st - siltstone, sst - sandstone, cgl - conglomerate, vf - very fine, f - fine, x-bed - cross-bedding, pgr - prograding; Locality abbreviations: G - Gressholmen, H - Hovedøya, L - Langøyene, R - Rambergøya.

<table>
<thead>
<tr>
<th>FA</th>
<th>Facies</th>
<th>Short description</th>
<th>Interpretation</th>
<th>Locality</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>A, B, C, D, P</td>
<td>st, vf sst</td>
<td>outer shelf</td>
<td>all</td>
</tr>
<tr>
<td>2</td>
<td>C, D, G, N</td>
<td>vf sst, f sst</td>
<td>middle shelf</td>
<td>all</td>
</tr>
<tr>
<td>3a</td>
<td>D, E, F, G</td>
<td>f sst, ripple lamination</td>
<td>inner shelf</td>
<td>all, H?</td>
</tr>
<tr>
<td>3b</td>
<td>D, G, P</td>
<td>f sst</td>
<td>inner to middle shelf</td>
<td>L</td>
</tr>
<tr>
<td>4a</td>
<td>I, J, E</td>
<td>clast supported cgl</td>
<td>fluvial, high energy</td>
<td>L, R</td>
</tr>
<tr>
<td>4b</td>
<td>I, L</td>
<td>trough x-bed sst and cgl</td>
<td>fluvial, middle to high energy</td>
<td>L</td>
</tr>
<tr>
<td>4c</td>
<td>C, I, L</td>
<td>sst, cut-and-fill structures</td>
<td>fluvial, middle energy</td>
<td>L, R</td>
</tr>
<tr>
<td>4d</td>
<td>M, N, (K)</td>
<td>boulders at channel base</td>
<td>fluvial channel margin</td>
<td>H</td>
</tr>
<tr>
<td>4e</td>
<td>J</td>
<td>clast supported cgl, prograding</td>
<td>bay-head delta</td>
<td>L</td>
</tr>
<tr>
<td>5</td>
<td>K, N</td>
<td>migrating dunes</td>
<td>near-shore</td>
<td>G, H</td>
</tr>
<tr>
<td>6</td>
<td>D, N, O</td>
<td>m sst</td>
<td>outer estuary</td>
<td>H</td>
</tr>
<tr>
<td>7</td>
<td>H</td>
<td>brown weathering sst</td>
<td>open shelf</td>
<td>all</td>
</tr>
</tbody>
</table>

**Interpretation**

The fine grain size (silt to very fine sand) indicates a low energy depositional environment. The fine-grained silt fraction in this facies association is thought to have been deposited well below fair weather wave base. Bioturbation indicates additionally a relatively low sedimentation rate and most likely oxic sea floor conditions. The hummocky cross-stratification and swaley structures further up in the succession indicate a storm origin of these layers. In this context the upwards thickening of the interbedded sand layers in facies association FA 1 can be interpreted as the development from low to middle energy conditions, related to shallowing and in increase in energy at the fair weather wave base, at which the sand was deposited (cf. Chapter 4.1).

The development of layer thickness throughout FA 1 marks a shift in depositional environment from the lower part of the offshore-transition to the upper part of the offshore-transition, closer to the fair weather wave base (cf. Fig. 4.1).

Middle sand-sized quartz grains have also been observed in the lower part of the succession on Langøyene, where they occur in beds with soft sediment deformation (facies P). Since the deformed beds occur at local stratigraphic levels in the succes-
sions, the deformation may be results of intense storm events, and not by late post-depositional tectonic processes (see discussion in facies P above).

According to the hummocky cross-stratification types, proposed by Dott and Bourgeois (1982), the appearance of FA 1 is likely to represent graded laminites, distal relative to the storm centre in relatively deep water (close to storm wave base).

6.2.2. Facies association 2 (FA 2) – middle shelf

Description

The lithologies in facies association FA 2 consist of very fine- to fine-grained sandstone, of facies C, D and G. Middle-grained sandstones of facies N may occur in some places interbedded. Both wave and current ripples and ripple lamination may occur in these layers, as well as parallel lamination in ca. 20 cm thick beds. Storm layers, often consisting of fine-grained sandstone, occur interbedded and may show low-angle cross-stratification or hummocky cross-stratification. Erosive bases may occur preferentially close to the conglomerate units (cf. Chapter 6.2.4).

Interpretation

The grain size of the background sediment has increased, indicating increase in energy by time due to shallowing by sediment accumulation, or fall in relative sea level. The
storm layers (facies D) in FA 2 appear to be very similar to those in FA 1. The storm-events appear to have continued during deposition of strata here defined to several facies associations. Small-scale cut-and-fill structures may be explained as the result of strong storm events that have produced channelized currents on the shelf, which in turn leave these sandstone channel-like structures. Wave ripple lamination (facies G) indicate that the sea bed must have been above fair weather wave base, at least for some time, likely of longer time than the duration of very strong storm events.

Facies association FA 2 most probably represents a relatively high energy depositional environment close to fair weather wave base on the middle to inner storm-dominated shelf.

### 6.2.3. Facies association 3 (FA 3) – inner shelf

**Description FA 3a**

Facies association FA 3a consists of 20 to 30 cm thick sandstone beds, showing numerous internal structures as prevalent current ripple- and frequent climbing ripple lamination (facies F), interbedded with sandstone beds showing hummocky cross-stratification and swaley structures (facies D). Thinner sandstone laminae (some centimetres thick) with wave ripples on the bedding surface may occur (facies G). Cross-stratified sandstones of facies E may occur locally. The grain size ranges from fine to middle sand. Facies association FA 3a occurs on Hovedøya and Rambergøya.
Description FA 3b

Facies association FA 3b has only been recorded on Langøyene. It combines the facies D, G and P. In contrast to FA 3a, almost no predominant ripple structures have been observed in this facies association. Grain size and thickness of the sandstone beds are slightly smaller compared to the other localities on Hovedøya and Rambergøya.

The sandstone beds in this locality are about 5 to 10 cm thick. Storm layers with hummocky cross-stratification and swaley structures occur in the succession and form often amalgamated beds. The grain size ranges from very fine to fine sand, and there were no gradations observed within individual sandstone beds. Very high up in the succession, wave ripples occur on one exposed bedding surface on top of a 4 m thick bedset characterized by soft sediment deformation structure (facies P). Apart from that, ripple lamination is very rare in this succession. Soft sediment deformation structures (facies P) are in this locality presumably developed as ball-and-pillow structures.

Interpretation

The sedimentary structures in FA 3a indicate storm-dominated depositional processes. The earlier deposited sand may have been reworked by storms, creating the internal sedimentary structures as low-angle cross-stratification and climbing ripple lamination. The sedimentation rate can be considered to have been relatively high due to the very abundant occurrence of climbing ripple lamination and also asymmetric current ripple lamination.

The sedimentary structures in facies association FA 3b are not as diverse as in FA 3a, but the character of bedding is very similar and is thought to reflect similar depositional processes. The sediments have been interpreted to represent rather distal storm-generated sediments: distal in terms of position of the storm centre. That would explain the similar appearance of the beds and the absence of structures indicating high sedimentation rates, such as climbing ripple lamination. Deformation structures may also be related to storm events. The massive and structureless feature of some of the thick beds with soft sediment deformation may point to a rapid deposition of the strata (e.g. by storms). This also implies that the sediments had not been affected by early-diagenetic cementation when the deformation was initiated.

The appearance of FA 3b is very similar to the mixed carbonate-siliciclastic facies association from a low-ramp setting in Jämtland, Sweden, also being a part of the epicontinental sea covering Baltica during Hirnantian times; the deposits of this area were interpreted to represent alternating fair-weather and storm-weather conditions (Dahlqvist and Calner, 2004). Amalgamated beds with internal hummocky cross-
stratification and occasional occurring contorted beds have been characterized by Dott and Bourgeois (1982) to indicate frequent storm events. Amalgamation and cross-stratification structures indicate a shoreface environment (Loi et al., 2010).

The depositional environment for facies association FA 3b is interpreted to have been middle to inner shelf position under intermediate to low energy conditions, with water depths close to fair weather wave base (FWWB), represented by the sections on Hovedøya and Rambergøya, and between FWWB and storm wave base (SWB) in the section on Langøyene, representing lower energy conditions, compared to the other studied sites. Wave ripples in all localities indicate that there must have been episodes where the deposition took place above FWWB.

The sediments of FA 3b have most likely been deposited under less energetic conditions than those of FA 3a. Wave ripples in the upper part of the succession on Langøyene (FA 3b) indicate that also the more distal parts were deposited in a position slightly above the FWWB for a very short period. An explanation for the locally occurring wave ripple lamination in FA 3a may be that storms reworked most of the sediments, which subsequently were reworked by waves hitting the sea bed during waning energy conditions.

6.2.4. Facies association 4a-c (FA 4a-c) – fluvial deposits

Description FA 4a

Facies association FA 4a consists of multi-laterally stacked cut-and-fill structures consisting of clast-supported conglomerate beds (facies I) and lensoid-shaped coarse sandstone (facies J). The stacking patterns may pass upwards into planar cross-bedded sandstone (facies E) and multi-lateral stacked cut-and-fill structures consisting of matrix-supported conglomerate beds (facies J).

Description FA 4b

Facies association FA 4b consists of trough cross-bedded middle- to coarse-grained sandstones (facies L), interbedded with locally occurring clast-supported conglomerate (facies I). FA 4a may be found below the trough cross-bedded sandstone beds. Dunes, climbing dunes and ooides occur at some places within facies L (Fig. 6.14B, C). Units of this facies are up to 5 m thick and are exposed on Langøyene.
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Description FA 4c

Facies association FA 4c consists of cross-bedded sandstone beds (facies I, L). The beds are generally stacked laterally, but in some places also minor, vertically stacking patterns are present (facies C). Clasts of coral fragments may occur at the base of these units but are very rare. The total thickness of beds in this facies association FA 4c is with some tens of centimetres relative thin, compared to FA 4a and FA 4b, which range in meter-scales.

Interpretation

Facies associations FA 4a, b and c are interpreted to represent the infill of fluvial channels characterized by the association of conglomerate and sandstone beds of 3D and 2D dunes (trough cross-bedded and planar cross-bedded beds, respectively), climbing dunes, and cut-and-fill structures. Facies association FA 4a appears to be deposited under relatively high energy conditions, involving transport and deposition of coarse, clastic material. The clasts are well rounded and elongated, which indicates abrasion of the clasts by flowing water. Deposits of FA 4c occur in localities with close lateral association to FA 4a deposits and may represent sites, where the fluvial current lost energy, due to declining energy of the overall fluvial stream, or to upstream changes in the fluvial channel system, as for example avulsion.

Deposits of FA 4b occur on Langøyene in the same level as FA 4a and FA 4b on Rambergøya. FA 4a occurs below FA 4b on Langøyene. This indicates that these facies associations are related in terms of decreasing flow energy from depositional areas of FA 4a to areas represented by beds of FA 4b. Lower-energy FA 4b beds appear thus to be lateral equivalent to the higher energy FA 4a deposits, as well as succeeding stratigraphically the FA 4a, as seen in the section on Langøyene. The depositional processes are assumed to have been the same during deposition of sediments of these two facies associations, only with differences in the size of the transported material, thus reflecting differences in the flow energy.

Similar deposits in West Africa have been interpreted to represent braided streams, infilling large depressions formed by glacio-related erosional phases (Ghienne, 2003). In this study, synsedimentary deformation structures have been described to be very abundant. However, the fluvial channel conglomerate units in the Oslo Region may represent something in between braided streams, passing upwards with decreasing equilibrium profile, into more meandering streams. This is expressed by large-scale cross-beddings in the finer-grained fluvial derived sandstones.
Figure 6.14: FA 4a-c of the Langøyene Formation: A second conglomerate unit in the eastern locality on Rambergøya, representing FA 4a; B trough cross-bedded sandstone, representing FA 4b with climbing dune structures, in the eastern profile on Langøyene, continuing upwards into cross-bedded, channelized sandstones, representing FA 4c; C multi-lateral stacked channel structures, representing FA 4a at the base, passing upwards into FA 4c. Note, that all photographs are facing southward, and therefore east is on the left hand side and west on the right hand side. Facies E: fine sandstone, planar cross-bedded sandstone; Facies I: clast-supported conglomerate; Facies J: lensoid-shaped sandstone; Facies L: matrix-supported conglomerate and trough cross-bedded sandstone.
6.2.5. Facies association 4d (FA 4d) – fluvial channel margin

Description

Facies association 4d consists of conglomerate with huge clasts, having diameters up to several meters (facies M, Figs. 6.7, 6.15). The clasts consist of trough cross-bedded calcareous sandstone, containing quartz grains and ooides (facies K). Locally, this facies association appears also as a conglomerate with small clasts in centimeter-scale (facies N).

Interpretation

As these conglomerate units consist of huge clasts up to the dimension of boulders, which likely have been too big to have been transported over long distances, it is here interpreted that they represent deposits close to fluvial channel margins. During fluvial incision and lateral undercutting of carbonate-cemented trough cross-bedded sandstone belonging to the underlying lower part of the Langøyene Formation, boulders and blocks may have collapsed along the margins and “fallen down” into the channels without being transported any long distance (see also discussion under facies M above).
6.2.6. Facies association 4e (FA 4e) – bayhead delta

Description

Facies association FA 4e is present at the western coast of Langøyene and consists of matrix-supported conglomerate units (facies J), which show prograding and horizontal bedded beds (Fig. 6.16). The grain size of these conglomerate beds are decreasing in direction of the progradation. Beyond the prograding beds, FA 4e continues into an approximately 50 cm thick bed of FA 4a. This facies association extends laterally over ca. 20 m (top of prograding beds to erosive end of horizontal beds). The prograding unit is up to 5 m thick at it’s maximum. The horizontal beds are thinner, reaching a thickness up to 3 m.

Interpretation

The prograding beds together with the horizontal beds have been interpreted to represent a small-scale bayhead delta, which was deposited after the maximum relative sea-level lowstand (cf. Figs. 4.4C, 4.6A, C). The lateral association with facies association FA 4a further east on Langøyene can be interpreted as representing tributary channels (Fig. 6.17, facies I), supplying sediment for the formation of the delta.

Bayhead deltas have been interpreted by different authors to form during sea-level stagnation (Allen and Posamentier, 1993) or transgression (e.g. Zaitlin et al., 1994). Bayhead deltas have often be assigned to the outer parts of incised valleys (Zaitlin et al., 1994), where they remain basically unreworked by fluvial or tidal processes (Nichol et al., 1997). Applied on the Langøyene Formation, this would lead to an interpretation of the formation of a bayhead delta during a sea-level lowstand.

6.2.7. Facies association 5 (FA 5) – near-shore

Description

Facies association 5 consists of both planar and trough cross-bedded sandstone. Planar cross-beds (facies K) show accretion surfaces that indicate westward transport directions. Trough cross-bedding shows no distinct transport direction, whereas parallel cross-bedding indicates a predominantly westward transport direction. Well rounded quartz grains and ooides were observed in the middle- to coarse-grained sandstone beds (facies K, N). Clasts consisting of coral fragments occur in some places at the base of the troughs. Individual beds are between 15 and 20 cm thick, graded and vertically stacked. Facies association FA 5 has been recorded at Hovedøya in the central southern and northern locality, as well as on Gressholmen (no logged section from the latter two
Figure 6.16: Facies association FA 4e at the southwestern coast of Langøyene; prograding beds and horizontal beddings, interpreted to represent foresets are indicated with white lines in the figure on the right hand side. The photograph was taken by Ivar Midtkandal.

localities). At the central southern locality on Hovedøya, this facies association occurs directly below an subaerial unconformity. The lateral extension of the individual beds is hard to determine since the outcrops are laterally very restricted. However, a lateral extend of up to ten or some tens of metres may be assumed due to little variation in bed thickness in the observed localities.

Interpretation

This facies association indicates a high energy environment where directional currents controlled transport and sedimentation and were powerful enough to carry middle to coarse sand grains and ooides. The high abundance of quartz grains and ooides suggests that this depositional environment may have been situated relatively close
to the source of these components, which may be the case close to a shoreline. The interbedded cross and planar cross-stratified units indicate changes in flow direction that may have been related to changing energy conditions. Facies association FA 5 is interpreted to represent migrating near-shore sand bars (Brenchley and Newall, 1977, 1980).

6.2.8. Facies association 6 (FA 6) – outer estuary

Description

Facies association FA 6 consists of vertical stacked, graded beds, which only occur in the southwestern locality on Hovedøya above the second major erosional unconformity. Internal structures occur as parallel lamination, current ripple lamination, HCS and swaley structures (facies D) and dewatering structures. Individual beds have erosive bases and the transition to the overlying bed is conformably. Interbedded with those laminated structures, coarse-grained sandstone lensoid beds occur, which contain clasts or abraded fossil fragments (facies N). In the uppermost part of this FA, two layers of coquina beds (facies O) were observed.

Interpretation

The deposition of coquina beds as well as relatively coarse, siliciclastic sediments indicate a high energy environment. In this specific setting, FA 6 is interpreted to represent a storm-dominated environment. During storm events, waves reworked previously deposited sediments. Coquina beds have been interpreted to represent widespread coastal storm-events (Radley and Barker, 2000).

6.2.9. Facies association 7 (FA 7) – open shelf

Description

Facies association FA 7 consists of the brown weathering sandstone (facies H, see description above, Fig. 6.17) at the top of the Langøyene Formation. It has been recorded at the top of all logged sections, laying unconformably on the karstic surface on top of facies K and conformable on top of sandstones of the infill succession.

Interpretation

Due to its extensive expansion and reworking by biota (see description and discussion of facies H above), the sandstone of this facies association has been interpreted to have
been deposited under relative sea-level stagnation and little new sediment influx.

**Figure 6.17.** Facies association FA 7 (facies H) at **A**: at the locality on eastern Langøyene, conformably on top of sandstones (facies F and D) and **B**: laying unconformably on top of the karstic surface at the locality on southern Hovedøya.
7 Log correlation and sequence stratigraphy

7.1. Log correlation

The sedimentary logs from the localities Rambergøya east and west, Langøyene east and west and the profile on the southwestern shore on Hovedøya were correlated by using major erosional unconformities (cf. Chapter 4.3.3). These surfaces have been traced throughout most of the study area and could therefore be applied for correlation purposes. The positions and abbreviations of the logged sections are given in Fig. 7.1.

Individual sequences are divided by subaerial unconformities. Each unconformity is characterized by a distinctive lithologic change from marine or estuarine sandstone (cf. Chapter 6.2, FA2, FA3) below, to fluviatile conglomerate and sandstone units (cf. Chapter 6.2, FA4) above, and represents thereby an abrupt basinward shift of facies. This property is typical for a sequence boundary (SB) defined as a subaerial unconformity as bounding surface of the deposition sequence (Mitchum et al., 1977; van Wagoner et al., 1988). The uppermost part of the Husbergøya Formation is shown in the correlated logs and has been included for the purpose of reconstructing of sea-level changes (Chapter 8.2).

The first datum plane for the log correlation was chosen to be the uppermost flooding surface between the brown weathering sandstone of the Langøyene Formation and the nodular limestone at the base of the Solvik Formation since this event is considered

Figure 7.1.: Schematic map of the positions of the island in the inner Oslofjorden, relative to each other. The position of the islands in the Oslo Region is shown in Fig. 3.1. The pink dots and box indicate the Log positions.
to represent a regional flooding event which most likely happened very fast.

The second datum plane is the flooding surface at the base of the Langøyene Formation. However, this surface may be diachronous and will therefore be used only where the datum plane at the base of the Solvik Formation is not exposed for correlation of the logs from Rambergøya east and west.

The correlation of the sedimentary logs from Rambergøya (Fig. 7.2) shows that both logs are very similar in terms of thickness of marine and estuarine facies development. However, the layers tend to be slightly thicker in the western locality. The oldest conglomerate unit (above SB1) is much thicker developed in the western locality, compared to the eastern one. The stratigraphic section above the second conglomerate unit in the eastern log is missing due to erosion at the southern shore of Rambergøya. The first and thin sequence (above SB1) may be explained by deeper erosion in the west.

The correlated logs from Langøyene (Fig. 7.3) show big differences in the thickness of their sequences. The first sequence is in the western locality much thinner than in the eastern locality. In contrast, the second sequence is much thicker in the western locality than in the eastern. The second sequence, consisting mainly of trough cross bedded sandstone in the eastern locality, is 5 m thicker here than in the west. In contrast, the marine/estuarine infill in the western locality is 15 m thick, whereas this section is 4 m thick in the eastern locality.

The correlation of the logs from Hovedøya, starting 26.5 m above the base of the Langøyene Formation, shows a distinct channel-shape of the erosional unconformity (Figs. 7.4, A.1). The positions of the logs and the infill of the incision are schematically shown in Figures 5.1 and 7.4. A karst surface on top of a calcareous sandstone bed (cf. Chapter 6.1, facies K) has been recorded in the southern locality on Hovedøya, indicating subaerial exposure of the respective unit.
Figure 7.2.: Simplified sedimentary logs from Rambergøya, showing in the western log the complete Husbergøya and Langøyene Formations and in the eastern log the uppermost meters of the Husbergøya Formation and the Langøyene Formation. The pink dots on the small map indicate the log positions.
Figure 7.3: Simplified sedimentary logs from Langøyene. The pink dots on the small map indicate the log positions. The legend for this figure is given in Fig. 7.2.
Figure 7.4.: Simplified sedimentary log correlation starting 26.5 m above the base of the Langøyene Formation on Hovedøya. The positions of the single logs is shown in the upper photograph. Facies association FA 6 represents high-energy cross-stratified sandstones. The exact boundary between FA 5 and Facies association FA 2 is speculative. Logs 1 to 5 are not exposed in the shaded area, the respective area has been extrapolated. A schematic sketch of the infill is also shown in figure 5.1. Note that the layers in the upper photograph are overturned, and the stratigraphic way up is downwards. The pink dot on the small map indicate the log positions of the southern locality on the right hand side, the pink box indicates the area, shown in the photograph. The legend for this figure is given in Fig. 7.2.
Figure 7.5.: Correlation panel of main logs from Langøyene, Rambergøya and Hovedøya, showing the uppermost meters of the Husbergøya Formation and the complete Langøyene Formation, as it is exposed in the considered localities. The legend for this figure is given in Fig. 7.2.
7.2. Sequence stratigraphy

A depositional sequence has been defined by Mitchum et al. (1977) to be a conformable succession, bounded at its top and base by unconformities and their correlative conformities. Three major erosional unconformities have been recorded in the Langøyene Formation (see Chapter 7.1). Following the concept by Mitchum et al. (1977), the Langøyene Formation has been divided into three sequences, described below. The sequences consist of sediments, filling in the accommodation space created during sea-level rise. Those sediments can be subdivided into systems tracts, which are defined to represent a linkage of contemporaneous depositional systems (Brown and Fisher, 1977). These systems are associated with a specific segment on the eustatic curve, resulting in lowstand, transgressive-, highstand and shelf-margin systems tracts (Posamentier et al., 1988). The lowstand systems tract (LST) is the stratigraphically oldest systems tract within a sequence, deposited during an interval of relative sea-level lowstand and subsequent rise in sea level (Myers and Milton, 1996). The transgressive systems tract (TST) develops during the maximum rate of sea-level rise and corresponding retrogradation of the shoreline (Posamentier et al., 1988; van Wagoner et al., 1988; Myers and Milton, 1996). The onset of the maximum rise in sea level is marked by a flooding surface (FS). Flooding surfaces are indicated with blue lines in the simplified logs and represent the lower boundary of the transgressive systems tracts in the Langøyene Formation (Figs. 7.2, 7.3, 7.4, 7.5 and Fig. 7.8).

7.2.1. Sequence boundaries on Rambergøya and Langøyene

The major erosional unconformities on Langøyene and Rambergøya can be traced in all four sections logged on these islands. They are marked with red lines in the simplified logs (Figs. 7.2, 7.3, 7.5). Their lithologic appearance in the localities on Langøyene is shown in Fig. 7.8. The relative erosional depths are shown schematically in Fig. 7.6. Due to the erosion of the uppermost part of the Langøyene Formation in the eastern locality on Rambergøya, the second datum plane (base Langøyene Formation) has been used for the estimation of relative differences in erosional depth.

The lowermost major erosional unconformity (SB1) is located 15.5 m above the base of the Langøyene Formation in the western locality on Rambergøya and 14 m in the eastern locality ($\Delta R_1 = 1.5$ m; Fig. 7.6). The respective erosional surfaces are recorded 26 and 29 m above the lower boundary of the Langøyene Formation (36 and 33.5 m below the datum plane at the lower boundary of the Solvik Formation) on Langøyene west and east, respectively ($\Delta L_1 = 2.5$ m; Figs. 7.8, 7.6).

The second major erosional unconformity (SB2) within the Langøyene Formation on
Rambergøya occurs at 32 and 29 m above the base of the formation in the eastern and western locality, respectively, giving a minimum difference in palaeotopography ($\Delta R_2$) of 3 m on a distance of approximately 400 m. In respect to the base of the Solvik Formation in the western locality on Rambergøya, the second major erosional unconformity is 22.75 m below this datum plane.

The respective unconformites (SB2) in the localities on the eastern and western shore of Langøyene are located 17.5 and 30 m below the base of the Solvik Formation (45 m and 32 m above the base of the Langøyene Formation). Based on this information, the second erosional phase has cut at least 12.5 m into the succession on Langøyene ($\Delta L_2$, Fig. 7.6). This value for the erosional depth representing palaeotopography in this area has to be seen as a minimum value. The locality on Rambergøya shows that the second major unconformity occurs within this minimum erosional depth of 12 m.

The third major erosional unconformity (SB3) on Rambergøya west is positioned 3.5 m below the datum plane at the base of the Solvik Formation. The local erosional relief between the eastern and western locality on Rambergøya cannot be given since the section in the eastern locality is eroded above the second conglomerate unit. At the eastern locality on Langøyene this surface lays 1.5 m, and 7 m below the lower boundary of the Solvik Formation in the western locality. These measurements give a minimum palaeotopographic erosional relief of 5.5 m for the third major erosional event.

**Figure 7.6:** Schema of the erosional depths and correlated palaeotopographic differences from Langøyene, Rambergøya and Hovedøya. $\Delta E_{1,2,3}$ give the total erosional depths for the first, second and third erosional event, respectively. The local palaeotopographic minimum depths are given in the respective column at their stratigraphic position, relative to the datum plane on top of the Langøyene Formation.
7.2.2. Sequence boundaries on Hovedøya

Two erosional unconformities (SB1, SB2) are observed on Hovedøya (Figs. 7.4, 7.6). The younger (SB2) of them has been formed by erosion into the underlying strata, deeper than the stratigraphic level of the older sequence boundary (SB1) (Figs. 7.7, A.2). The older unconformity occurs as an erosional surface cut to a shallow depth, presently located 31 m above the base of the Langøyene Formation. The oldest sequence boundary in the Langøyene Formation has been identified by a layer of conglomerate that is exposed directly above the erosional unconformity (Figs. 7.7, A.2). This conglomerate unit is overlain by sandstone beds with parallel stratification. Those sandstone beds also are partial reworked in a lensoid body of deformed shelf-sediments. The deformation of the respective beds has most likely taken place during the second erosional event.

The second incision cuts at least 10 m into underlying sediments. The subaerial unconformity on the southwesternmost shore of Hovedøya, where no conglomerate unit is present, may have been part of an interfluve area, representing rapid facies changes over a distance of 200 m.

The area between the parallel log 8 (Fig. 7.4) and the locality on southern Hovedøya, interpreted to be part of an interfluve area, was not exposed due to tectonic faulting and overgrowing vegetation. Since the transition from fluvial margin facies towards the interfluve area is not exposed, the value for the local palaeotopographic change has to be seen as a minimum value and might be larger (Fig. 7.6). At the observation point Hø (Fig. 7.1), in the eastern part of the assumed interfluve area, a conglomerate unit (cf. Chapter 6.1, facies M) crops out at the shore, indicating that the area in between was actually exposed between two incisions.

7.2.3. Correlation of the sequence boundaries and consequences for the sedimentation model

The third sequence boundary (SB3) on Rambergøya and Langøyene has been shown to be associated with the weakest sea-level fluctuation compared to SB1 and SB2, since its deepest position is located 5 m in the underlying strata and between Rw and Lø. SB1 and SB2 are represented by deeper incisions on both Langøyene and Rambergøya compared to that of SB3.

Based on the facies distribution and the significance of the sequence boundaries, the two major erosional unconformities on Hovedøya have been interpreted to represent SB1 and SB2. This correlation influences the estimation for the palaeotopographic differences during the time of the erosional events to a minimum of $\Delta E_1$ of 22 m and $\Delta E_2$ of...
The previously described three major erosional unconformities, defined as sequence boundaries, are interpreted to have been formed by changes in relative sea level (see also Chapter 8.2). The actual erosional depth and accompanied palaeotopography in this area may have been much deeper since it is very unlikely that the logged sections represent the deepest and shallowest localities, respectively.

7.2.4. Sequence one – S1

The early infill above the first sequence boundary consists of fluvial conglomerate beds, defined to be facies association FA 4 (cf. Chapter 6.2). Three backstepping trends have been recorded in the lowermost conglomerate unit on Langøyene (Figs. 7.8, 7.5).
Figure 7.8.: Erosional unconformities on Langøyene, interpreted as sequence boundaries (SB), with their stratigraphic position indicated on the simplified log from western Langøyene: SB1 in the western locality on Langøyene; SB2 in the eastern locality on Langøyene, marked in the respective unit in the profile from western Langøyene; SB3 in the western locality on Langøyene, followed overlain by a conglomerate unit, interpreted to represent a bayhead delta. Note that all figures are oriented southward and east is on the left hand side and west on the right hand side, respectively.
CHAPTER 7. LOG CORRELATION AND SEQUENCE STRATIGRAPHY

These three phases may be subdivided into two main phases, which can also be traced in the localities on Rambergøya (Fig. 7.5). The early infill on Rambergøya consists of conglomerate beds interpreted to represent facies association FA 4a and FA 4c in the western and eastern locality, respectively (cf. Chapter 6.2). On Langøyene, the early infill-units consist of a combination of facies association FA 4a and FA 4c, whereas FA 4a represents the lower infill of each phase, passing upwards into FA 4c (Fig. 6.14). The first early infill on Hovedøya was mainly eroded during the formation of the second subaerial unconformity. However, a thin bed of conglomerate is exposed above the first sequence boundary (Fig. 7.7). This layer represents an erosional remnant of the early infill sediments of the incised valley system on Hovedøya. This lowermost fluvial infill may be referred to as the lowstand systems tract (LST) within the first sequence. The LST is bounded at the top by a flooding surface, marking the onset of the transgressional phase.

The sediments deposited during ongoing transgression consist of sandstone beds, showing numerous internal structures as (climbing) ripple lamination and hummocky cross-stratification (cf. Chapter 6.2, FA 2, FA 3). They indicate the transition from fluvial sediments to estuarine to marine sediments. This succession may be referred to as the transgressional systems tract (TST).

A karstic surface, interpreted to represent SB2 in the southern locality on Hovedøya (Hs), has developed on top of cross-bedded, ooid-containing, calcareous sandstones (facies K, Figs. 3.6, 7.4). Given that the cross-bedded sandstones (facies K) are laying on top of facies associations that indicate middle shelf depositional environments, they can be interpreted to represent the termination of a regressive trend, ending with the subaerial exposure of this area (Fig. 7.4 Hovedøya south). Giving the assumption that the underlying strata of the carbonate sandstone beds (facies K) represent a regressive trend, then it is likely to record a transition from FA 2 (exposed) into FA 3 and FA 6 (not exposed), which represent near-shore facies and would reasonably be deposited on top of middle and inner shelf sediments. However, a transition from FA 2 to FA 3 to FA 6 is not documented by exposures.

7.2.5. Sequence two – S2

The second (middle) sequence in the Langøyene Formation starts at the second sequence boundary (SB2). The early infill succession after the second erosive phase shows two main back-stepping trends recorded in the localities on Langøyene and eastern Rambergøya (Fig. 7.5).

The lithologies of the early transgressive infill consist of clast-supported conglomerate beds (FA 4a) or corresponding trough cross-stratified sandstone beds (FA 4b), pass-
ing upwards into planar cross-bedded sandstones (cf. Chapter 6.2.4, Figs. 6.14, 7.8) in the localities on Rambergøya and Langøyene. The corresponding infill succession on Hovedøya consists of the conglomerate unit composed of huge blocks. This unit is rather thin (two to three meters, depending on the block size) and lacks pebble-infill (FA 4d, cf. Chapter 6.2.5). No internal flooding surfaces were recorded in this unit on Hovedøya. These fluvial conglomerate units, deposited during relative sea-level lowstand, are referred to as the lowstand systems tract within the second sequence.

The onset of the transgression is marked by a flooding surface on top of the conglomerate units of the LST. The subsequently accumulated sediments have been interpreted to consist of FA 3 (cf. Chapter 6.2) and represent the transgressive systems tract.

### 7.2.6. Sequence three – S3

The third sequence boundary (SB3), is overlain by a conglomerate unit, varying in thickness from 0.3 to 5.5 m in the localities on Langøyene and western Rambergøya, respectively. This part of the section is absent due to erosion in the locality on the eastern side of Rambergøya. The third conglomerate unit is developed as a thin bed (30 to 50 cm) in the localities on western Rambergøya and eastern Langøyene, which was interpreted to represent the infill of fluvial channel deposits (FA 4a). The same unit is five meters thick in the western locality on Langøyene, showing prograding foresets, which has been interpreted to represent a bayhead delta (cf. Chapter 6.2, FA 4e). The fluvial conglomerate units (FA 4a, FA 4e) may be referred to as the lowstand systems tract in the upper sequence within the Langøyene Formation. The LST is overlain by a thin unit of sandstone beds, interpreted to represent FA 3 (cf. Chapter 6.2). These sandstone units represent the sedimentation during the ongoing transgression (transgressive systems tract).
8 Depositional history and sedimentary model

8.1. Depositional history of the Langøyene Formation

The depositional model of the Langøyene Formation obtained from the present study is shown in Figures 8.1 and 8.2. Figure 8.1A, C and E show the stages of relative sea-level highstand. Figures 8.1B, D and 8.2F show the three different stages of incision with fast lateral facies changes caused by emerged areas during lowstand of the relative sea level.

During stage B (Fig. 8.1B), the first incision occurred, due to a significant fall in sea level. Whereas the depositional environment on Hovedøya may have been characterized by sediment-bypass, the incisions on Langøyene and Rambergøya were succeeded by deposition of fluvial conglomerate beds. In stage C (Fig. 8.1C), the relative sea level rose and the first incised valley system was subsequently filled.

Stage D (Fig. 8.1D) shows the second significant fall in sea level, which caused the incision of extensive valleys in which clast-supported conglomerate units were deposited on Rambergøya and Langøyene. In the model, these conglomerate units are supposed to have been deposited simultaneously with the conglomerate units on Hovedøya. This assumption is based on the extension of the incision events, of which the first and second appear to have had the largest effects. Stage E (Fig. 8.2E) shows the next relative sea-level highstand, which was of lesser magnitude compared to stage C. The deposited sediments are interpreted to represent middle to inner shelf facies during this stage, close to or even slightly above FWWB (cf. Chapter 6.2.3, FA 3).

The relative sea level fell during stage F, but less compared to the previous two lowstand-stages (cf. Fig. 8.1B and D). A small bayhead delta was formed in the western locality on Langøyene, indicating a progradation towards southwest, forming a several meters thick conglomerate unit in this particulate locality. In the other localities both on Langøyene and Rambergøya, the incisions were less prominent and the early transgressive infills are present as some tens of cm-thick conglomerate units, which are interpreted to represent fluviat tributary channels for the bayhead delta. Due to the rather long distance between the localities, it is possible that the thin conglomerate units on Langøyene and Rambergøya belong to two different delta-systems, which may have existed parallel to each other. However, since only one clear delta has been found in this study, a single bayhead delta is shown in the model. The succession on
Figure 8.1: Model for the deposition of the Langøyene Formation shown in order A - D. The considered localities from logs or observations are indicated in G (Fig. 8.2). L, R and H on the single blocks indicate the localities Langøyene, Rambergøya and Hovedøya.
Figure 8.2.: Model for the deposition of the Langøyene Formation shown in order E - G. The considered localities from logs or observations are indicated in G, L, R and H on the single blocks indicate the localities Langøyene, Rambergoya and Hovedøya. The transparent drawn water areas in F indicate uncertainties of how drainage systems were present in the area of Hovedøya. The sketches and the relative position of the outcrops are not to scale but connections of fluvial channels between the localities may have been accomplished through meandering systems. The scale bars give an approximation of the extension of the study area during deposition.
Hovedøya does not show any evidence for a third erosional phase or subaerial exposure. Therefore stage F (Fig. 8.2) is shown as a submerged area. There is little evidence for whether there are tributaries for possible deltas on Hovedøya since lateral extending outcrops are disturbed by faults or overgrown by vegetation. It is possible that tributaries started northeastward of the deltas, even though this is very speculative (transparent water areas in Fig. 8.2F).

Stage G (Fig. 8.2G) shows the flooding of the whole study area, the formation of a regionally extensive flooding surface and the deposition of the brownish weathering sandstone that lies on top of both relatively continuous sedimentary successions and previously subaerially exposed carbonate platforms.

Evidence for the occurrence of interfluve areas was found on Hovedøya (Figs. 8.1B, D and 8.2E, F) where a karstic surface on top of an oolithic limestone has been interpreted to have formed in an interfluve area, simultaneously to the development of the incised valleys close by. This surface is the most significant indication for subaerial exposure, which can be found in the central southern shore-locality on Hovedøya (Fig. 3.6).
Figure 8.3.: A: hypothetical model of the deposition of the Langøyene Formation if there have been no incisions; B: model of the formation of the Langøyene Formation with glacioeustatic sea-level changes causing incisions. On the right hand side there are indications for the measurements of relative sea-level changes relative to the datum plane.
CHAPTER 8. DEPOSITIONAL HISTORY AND SEDIMENTARY MODEL

8.2. Implications for sea-level fluctuations of the Hirnantian in the central Oslo Region

The depth of fair and storm weather wave bases are strongly dependent on the period and height of the waves (Snedden et al., 1988). Therefore, depths of fair weather and storm wave base vary considerably, depending on the palaeogeographic position and exposure to dominating wind direction.

The fair weather wave base is most likely between 7 and 15 m depth if very fine sand has been transported by wave activity (Snedden et al., 1988; Plint, 2010). Embry and Klovan (1972) presented a depth of 9 m for the fair weather wave base, based on studies on coral- and stromatoporid-reefs. A study from a high-latitude, storm-dominated shelf setting during the Ordovician by Loi et al. (2010) set the fair weather wave base at a depth of 30 m. In this study, a 15 m weather-depth is estimated as the minimum depth for sediments deposited close to fair weather wave base.

The studies by Embry and Klovan (1972) and Loi et al. (2010) gave estimations for the storm wave base at depths of 21 and 120 m, respectively. This huge difference underlines the importance of the position of the particular study area in terms of palaeogeography and palaeo-oceanography. Studies by Allison and Wells (2006) showed that annual storms may rework sediments down to a depth of 40 m. The grain-size of the deposits, wave height and bedload transport vary during storm events (Plint, 2010). Snedden et al. (1988) suggested a water depth of 30 m for the transport of very fine sand during moderate storms.

It has been shown in this study that the Langøyene Formation in the central Oslo Region was deposited on a storm-dominated shelf, where fine- to middle-grained sand prevailed. Brenchley et al. (1979) suggested that the formation of the storm-layers was due to hurricanes, meaning that the water depth for the storm wave base even may have been deeper than the figures presented by Allison and Wells (2006), given that these events are larger than annual storms. The sea-level curve presented here is based on the assumption of the storm wave base at a depth of 35 to 45 m.

The reconstruction of the sea-level fluctuations (Fig. 8.4) is based on changes relative to the datum plane at the base of the Solvik Formation (cf. Fig. 8.3), suggesting that the karstic surfaces on top of the Langøyene Formation, beneath the transgressive surface at the base of the Solvik Formation, represents an emerged interfluve area.

The schematic development of the succession, as if there were no sea-level fluctuations, is shown in Fig. 8.3A. The shallowing trend can easily be explained by the successive infill of accommodation space. The development of the infill with temporarily
Figure 8.4.: Relative sea-level curve for the central Oslo Region during the Late Ordovician. The relative sea-level is given relative to the datum plane at the top of the Langøyene Formation. The assumed depths of the fair weather wave base (FWWB) and storm wave base (SWB) in this setting are indicated by the dark grey shaded area. S - Silurian, L - Llandovery, R - Rhuddanian. The lower boundary in the Upper Ordovician succession is uncertain and may lie in the Husbergeoya Formation. A-F represent the sea level, as they are shown in Figs. 8.1 and 8.2.

- The major sea-level lowstands (Figs. 8.1, 8.2 and 8.4B, D, F) are given relative to the datum plane as their stratigraphic distance, which gives in addition to the actual sea level above the datum plane, the total change in sea level (cf. Fig. 8.3B).
- The glacioeustatic induced sea-level fluctuation in the Oslo Region during the Hirnantian has been at least 60 to 70 m, according to the present data and the model of palaeowater depth prior to fall in sea level (Fig. 8.4). The following relative sea-level rise advanced in at least two steps in the early infill of the first sequence. The subsequent rise in relative sea level (Fig. 8.4C) remained most likely below the initial sea-level highstand (Fig. 8.4A) and was followed by a new drop in sea level (Fig. 8.4D). This second drop in relative sea level during the Hirnantian was less significant, but the minimum amplitude has been estimated to have reached 30 m. Stage E is charac-
terized by a relative constant sea level, followed by a short drop at its top (stage F), before the sea level rose significantly towards the Early Silurian. The sea level before and after the fluctuations have been assumed to be similar.

Since the used and visualised values in Fig. 8.4 are assumptions for the minimum incisions in the reference section for this study on Rambergøya west, the actual sea-level fluctuation has likely been larger.
Carbon isotope and XRF analysis

9.1. Carbon isotope analysis

9.1.1. Background

The carbon isotope record $\delta^{13}C$ is of great value not only due to its importance for stratigraphic correlation but also its potential to assist in investigations of the Earth’s climate, evolution of its biota and carbon dioxide levels (Saltzman and Thomas, 2012). However, there is no consensus model for explaining changes in the isotopic composition and the application as a chronostratigraphic tool has to be treated carefully (Ghiennie et al., 2014). All environmental implications are dependent on our understanding of the major changes in the carbon cycle, which may be influenced by e.g. carbonate-producing microorganisms in the oceans, abundance and evolution of plants and their photosynthetic mechanisms (Brenchley et al., 1994; Marshall et al., 1997; Saltzman and Thomas, 2012). Variations in $^{13}C/^{12}C$ in dissolved inorganic carbon (DIC) have been recorded through studies on marine carbonate rocks ($\delta^{13}C_{carb}$) globally through time (Saltzman and Thomas, 2012). Studies on isotopic compositions of organic carbon ($\delta^{13}C_{org}$) in rocks as black shales have been carried out and show a similar positive carbon isotope excursion for the Hirnantian (Marshall et al., 1997).

Positive stable isotope excursions for both oxygen $\delta^{18}O$ and carbon $\delta^{13}C$ have been reported from the Hirnantian, followed by a rapid fall of these values in the early Silurian (e.g. Brenchley et al., 1994; Marshall et al., 1997; Kump et al., 1999). This HICE (Hirnantian Isotope Curve Excursion) is of high regional chronostratigraphic value, especially for shallow-water carbonates from Baltica and eastern Laurentia (Brenchley et al., 2003; Delabroye and Vecoli, 2010; Kröger et al., 2015). This excursion might even be a global chronostratigraphic signal due to similar shapes and magnitudes of carbon isotope excursions in the Baltic region (Sweden, Estonia, Latvia), Laurentia (Anticosti Island, Canada) and close to the margin of Gondwana (Argentina) (e.g. Marshall et al., 1997; Kaljo et al., 2001; Brenchley et al., 2003; Hints et al., 2010).

The amount of inorganic carbon in the oceans and atmosphere varies on time scales of several thousand years as a result of imbalances between carbon inputs from weathering, metamorphism and volcanism, organic matter, derived from sediment burial and carbonate minerals (Kump and Arthur, 1999).
CHAPTER 9. CARBON ISOTOPE AND XRF ANALYSIS

A two-steps positive excursion in continuous strata of epicontinental seas for the Hirnantian was recorded by Brenchley et al. (1994); Kaljo et al. (2001) and Delabroye and Vecoli (2010) from Sweden and Baltica, Estonia and China, respectively. Investigations on whether the first or second peak is recorded have to be done in studies considering uncontinuous successions (Delabroye and Vecoli, 2010). Single peaks may also represent disjointed parts of several peaks and potential hiatuses have to be treated carefully (Ghienne et al., 2014).

More important than the actual values of δ¹³C are the trends in the curve, since the numerical values are dependent on e.g. study methods and bathymetry (Kaljo et al., 2004a). Melchin and Holmden (2006) argued that regional variations of the relative sea level on epicontinental shelves may have had a high influence on carbon cycling and consequently on δ¹³C isotopes, especially if there is a high increase in carbonate platform weathering during a relative sea level lowstand as it may have been during the Ordovician. The observed δ¹³C isotope excursions would then be a regional restricted event, predominantly occurring in restricted, shallow-water environments (Melchin and Holmden, 2006).

If an independent age control is available, isotopic excursions larger than 1-2‰ may be applied confidently (Saltzman and Thomas, 2012).

Kump and Arthur (1999) gave examples on how and to which extent δ¹³C isotopes may be influenced. The two strongest causes for positive inorganic carbon isotope excursion may be: a) increased burial of organic carbon, or a reduction in organic carbon weathering would lead to a fall in atmospheric $p\text{CO}_2$, which would in turn cause a positive isotope curve extinction in both organic and inorganic carbon; b) increased proportion of carbonate weathering, relative to organic and silicate weathering, may cause positive shifts of the isotopic composition in the ocean. Marshall et al. (1997); Kump et al. (1999) agreed, though they had different hypothesis of what caused the isotope excursion, in the case the excursions are coupled to glacial events (Kaljo et al., 2004a). Finney et al. (1999) argued that the carbon isotope excursion is likely to be the result of glaciation and associated sea level fall.

Cold water usually has higher δ¹³C values in dissolved inorganic carbon (DIC) than warm water (Saltzman and Thomas, 2012). Photosynthesis causes depletion in $^{12}\text{C}$ in DIC in surface water, since it is restricted to the photic zone. Therefore δ¹³C values from water below this photic zone are lower than those in the water close to the surface (Saltzman and Thomas, 2012).

Brachiopods and marine cements from Sweden and Estonia show δ¹³C values of +3‰ in pre-Hirnantian times, changing to values $>$+4‰ or even as high as +7‰ (Marshall et al., 1997) in Hirnantian times. Meteoric diagenesis may alter δ¹³C towards more
negative values (Saltzman and Thomas, 2012).

If the shifts in $\delta^{13}$C are caused by increased weathering of carbonates (Kump et al., 1999) that have been exposed due to glacioeustatic sea level fall, the two main peaks in the $\delta^{13}$C curve may be correlated to two main phases of glaciation, which were described e.g. by Ghienne (2003).

One has to keep in mind that in different studies there has been measured on different carbon sources, as organic carbon ($\delta^{13}$C$_{org}$), which is often used in studies carried out in southern China, or inorganic carbon (DIC, $\delta^{13}$C$_{carb}$), as used in studies in eastern Laurentia and Baltica (Delabroye and Vecoli, 2010). Delabroye and Vecoli (2010) compared studies from Laurentia, southern China and the Canadian Arctic (Finney et al., 1999; Chen et al., 2006; Melchin and Holmden, 2006) and showed that shallow-water carbonates appear to produce stable $\delta^{13}$C$_{carb}$ signals, which are assumed to be isochronous in epicontinental seas (Delabroye and Vecoli, 2010). On the other hand, $\delta^{13}$C$_{org}$ signals for the Hirnantian show to be highly variable (Delabroye and Vecoli, 2010). In contrast, deeper-water shaly-strata produce consistent $\delta^{13}$C$_{org}$, but variable $\delta^{13}$C$_{carb}$ curves (Delabroye and Vecoli, 2010). The $\delta^{13}$C peaks for ocean basins and epicontinental seas have been demonstrated to be asynchronous due to restricted water exchange, and therefore delayed chemical responses to chemical variations take place in this environment (Melchin and Holmden, 2006).

Photosynthesis of both marine phytoplankton and land plants may influence $\delta^{13}$C values through atmospheric $p$CO$_2$ or the burial of carbon, beyond the previously named environmental and glacioeustatic effects. Plants can be categorised as C3, C4 or CAM, according to their photosynthesis pathways (Ehleringer et al., 1997). Recent plants mainly use the C3 photosynthetic pathway (Maslin and Thomas, 2003; Saltzman and Thomas, 2012). It has been shown that plants with the photosynthetic pathway C4 tend to become dominant over C3 plants with decreasing humidity at low $p$CO$_2$ values (Ehleringer et al., 1997). As described by Desrochers et al. (2010), the climate during the Late Ordovician may have been relatively arid. C4 plants, which have a tendency to produce relative higher $\delta^{13}$C values (average of -13 %) compared to C3 plants (average of -27 %) (Ehleringer et al., 1997; Maslin and Thomas, 2003). However, land plant fossils are recognised about 50 Ma after the appearance of land plant spores in middle Ordovician times (Gray, 1993; Kenrick and Crane, 1997). The poor fossil record and preservation potential of lower eukaryotes may have hidden their evolution through the Early Proterozoic when O$_2$ levels increased (Knoll, 1992) and plants may have been established on land before the fossil records at the latest from the Silurian and Devonian (Kenrick and Crane, 1997).
9.1.2. Results

The succession through the Husbergøya and Langøyene formations on Rambergøya has been sampled every second meter for analysis of $\delta^{13}$C. The results of the $\delta^{13}$C bulk-measurements, representing micritic carbonate cement of the rock samples analysed, are shown in Figure 9.1 and Table H.1. The $\delta^{13}$C value from 16 m above the lower boundary of the Langøyene Formation at Rambergøya was excluded from the diagram, since the measurements were acquired from a coral fragment. The analysed carbon from the coral may differ in $\delta^{13}$C from the $\delta^{13}$C bulk samples, even though it belongs chemically to inorganic carbon.

The curve starts with values around zero within the lower Husbergøya Formation and the values start to increase after 10 m towards the end of the formation. This has been interpreted to be the onset of the HICE. The $\delta^{13}$C curve in the lower part of the Langøyene Formation appears as a plateau with values ranging from 2.5 to 4.5‰. The values decrease above the first conglomerate unit, throughout the transgressive phase, until the lower boundary of the second conglomerate unit (sequence boundary), but still have their local minimum at values about 1‰.

A second, clear peak can be found above the second sequence boundary, within the conglomerate unit. This peak is likewise the maximum peak for the whole measured succession with a value of 5‰. It is followed by a steep decrease of the $\delta^{13}$C values to even negative (-1‰) values. The isotope values remain at low levels around zero until the third sequence boundary, where they fall again into negative values and increase towards the lowermost layers of the Solvik Formation.

The first plateau of $\delta^{13}$C values in the lower part of the Langøyene Formation may include the first main peak of the HICE (Hirnantian Isotope Curve Excursion), whereas a clear outstanding peak is missing. The maximum of this lower part of the excursion lies above the first sequence boundary, within the early infill succession. It can be concluded that the HICE occurs throughout a succession of 62 m in the western locality on Rambergøya.

The $\delta^{13}$C curve obtained from samples taken in Langøyene shows generally lower values than the curve represented by the samples from Rambergøya. The sampling intervals in the Langøyene locality were inconsistent. Due to this fact, it is more difficult to see clear trends in the development of the carbon isotope curve.

An increase towards the top of the Husbergøya Formation is visible, and in the lowermost part of the Langøyene Formation the $\delta^{13}$C values appear to be relatively stable around 4.5‰. Likewise, as in the curve from Rambergøya, this plateau is followed by a decrease to values around 1.5‰. Above the first sequence boundary the measured
Figure 9.1: To the left, detailed $\delta^{13}$C\textsubscript{carb} curve (in ‰ relative to the Vienna Peedee belemnite (VPDB)) of the Husbergøyøya and Langøyene formations in western Rambergøyøya with spacing of 2 m between the samples. For comparison, on the right hand side, the stratigraphic section on Langøyene with larger sample spacing.
sample shows again a relatively high value of 4%, and within the lowermost infill succession above the second sequence boundary, the $\delta^{13}$C values appear to be either very variable or show a rapid increase of 1‰ from 1.5 to 2.5‰. The values of $\delta^{13}$C seem to decrease relatively continuously towards the Solvik Formation without any outstanding peaks.

### 9.1.3. Comparison of the carbon isotope curves

The comparison of both measured $\delta^{13}$C curves in this study is not trivial due to different spacing between the samples. It must also be considered that there are very thick layers affected by soft sediment deformation in the succession in the Langøyene locality. This makes it difficult to obtain high-resolution measurements in the upper part of the Langøyene Formation from this locality. It seems, however, that the peaks and plateaus coincide with the occurrence of the first and the second conglomerate units, directly above the respective sequence boundaries. Due to deep erosional unconformities it is possible that strata with major peaks of the HICE have been eroded and the peaks appear to be related to the early infill successions, whereas possible original peaks may have been below the actual sequence boundary. However, in this study, the peaks are recorded above the two major sequence boundaries, respectively, and are interpreted to have been there initially.

It is rather difficult and maybe even speculative to locate peaks of the $\delta^{13}$C curve from the Langøyene samples. However, there are still two peaks that may possibly correlate with the peaks from Rambergøya (Fig. 9.1): The first one at the base of the first conglomerate unit (about 27 m above the base of the Langøyene Formation in the $\delta^{13}$C curve from the Langøyene locality), directly above the first sequence boundary, and the second one may be found in the abrupt increase of values above the second sequence boundary in the respective conglomerate unit (about 32 m, respectively). The plateau in the curve in the lowermost part of the Langøyene Formation may correlate with the similar plateau in the $\delta^{13}$C curve from the Rambergøya locality. However, since the $\delta^{13}$C values are very high in the curve from Langøyene, this plateau might correspond to the first plateau in the curve from Rambergøya. As previously commented, these correlations are very uncertain due to the lack of more narrowly taken samples and isotope data from this locality.
9.2. XRF analysis

As described in Chapter 5.2.3, the content of specific elements (Ti, Zr, Rb, Ca, Sr) can be used as a proxy for changes in the depositional environment. The results of these are given in Tables I.1 and I.2. Sample numbers 1 to 10 represent the Husbergøya Formation (cf. Chapter 3.2.1) and numbers 11 to 37 the Langøyene Formation. Sample 38 was taken in the lowermost nodular limestone in the Solvik Formation (cf. Chapter 3.2.3).

The content of titanium (Fig. 9.2) shows a peak ca. 10 m above the base of the Husbergøya Formation and decreases upwards almost continuously throughout the Husbergøya and Langøyene Formation. The peak at 10 m above the base of the Langøyene Formation (sample 15) was measured on rock samples close to storm-deposited layers. This development of the Ti content in the sediments indicates a coarsening upwards throughout the Langøyene Formation.

The content of zirconium (Fig. 9.2) shows a decreasing trend between the lower boundary of the Langøyene Formation and up to the first erosional unconformity (first conglomerate unit). Between the first and the second unconformities, the Zr values remain relatively high and start decreasing above the second unconformity throughout the remaining part of the Langøyene Formation. A marked increase in the Zr content is present ca. 10 m above the base of the Husbergøya Formation, coinciding with similar peaks in the curves for Ti and Rb. The development of the Zr curve indicates a relatively low content of fine grain size material in the lower Langøyene Formation, increasing towards the middle part (20 m to 34 m, corresponding to sample numbers 20 to 27), where the portion of coarser grains (sand) remains relatively high. Coarser grain sizes are more likely to consist of more immature mineral compositions (containing coarser quartz grains and feldspars), whereas finer components are likely to consist of more mature material (clay minerals, carbonate muds with little siliciclastic content).

The content of rubidium (Fig. 9.2) shows a decreasing trend throughout the whole Langøyene Formation, starting already below in the Husbergøya Formation. The maximum peak of the Rb content in the sediments occurs ca. 10 m above the base of the Husbergøya Formation. Between the first and second erosional unconformity, the Rb values remain at a very low level, contrary to the Zr values in the same part of the formation. A minor positive peak occurs above the second unconformity. The values above this minor peak decrease towards the Solvik Formation. The decrease in Rb through the lower to middle part of the Langøyene Formation can be interpreted by the recorded decrease in clay content, which becomes higher again towards the uppermost...
part of the formation.

![Figure 9.2:](image)

The often opposed values from Rb (overall decrease, lower values) and Zr (relatively stable at high values in the middle of the formation) are interpreted to reflect the overall coarsening upwards trend in the Langøyene Formation. This observation coincides as well with the clearly decreasing Ti values, indicating an increase in siliciclastic content.

The curves for calcium and strontium (Fig. 9.3) look overall similar. The highest content of both elements can be found in the lowermost part of the Langøyene Formation, increasing towards the first erosional unconformity. The maximum positive peak can be observed directly above the first unconformity. Between the first and second unconformity, Ca and Sr show a clearly decreasing trend. The next distinct positive peak occurs at the second unconformity, followed by a gradual decrease, similar to the development after the first unconformity. Towards the top of the Langøyene Formation the values for Ca and Sr slightly increase.

The XRF-curves for calcium and strontium indicate an overall decrease in carbonate content of the sediments, with moderate values in the middle of the Langøyene Formation. The high Ca values in the middle coincide with high Zr values, which were interpreted to occur due to coarser grain size, representing more immature sediment compositions. The general developments of the XRF-curves in figures 9.2 and
Figure 9.3: Calcium (Ca) and strontium (Sr) development throughout the stratigraphic section on western Rambergoya. The content of the respective element is plotted against the x-axis, the sample number is plotted against the y-axis. The sampling distance is 2 m.

9.3 reflect and confirm the overall upwards coarsening of the sedimentary succession, which has been recorded by sedimentary logging. The decreasing carbonate content in the sediments with interim excursions of Ca, Sr and Zr can be interpreted to reflect phases with higher siliciclastic and therefore less mature sedimentary input.
10 Discussion

10.1. Sedimentology and sea-level fluctuations

The sequence stratigraphic analysis of the sedimentary successions representing the Langøyene Formation on the innermost islands in the Oslofjorden has shown that there have been developed two major and a third, less extensive, unconformities that have been defined as sequence boundaries. The major erosional unconformities cut into a succession deposited in an epicontinental carbonate ramp-setting in the east of the developing Caledonian orogen (cf. Chapter 3.1, Fig. 3.3). The sequence boundaries have been used for correlation of logs obtained of the sedimentary succession on Hovedøya, Rambergøya and Langøyene. Additional observations from Hovedøya and Gressholmen have also been included in this study.

The erosional unconformities in the Langøyene Formation have long been described and many hypotheses for their formation have been suggested, as described in Chapter 3.2.2. The interpretation that the conglomerate units represent infills of tidal channels (Brenchley and Newall, 1975, 1980; Brenchley et al., 1979; Worsley and Nakrem, 2008) is unlikely to be valid for following reasons: (1) no signals of tidal forces (double mud drapes, flaser bedding, structures indicating bidirectional flows (Dalrymple, 2010)) have been recorded in the Langøyene Formation; (2) the recorded incision into the epicontinental sea sediments of the Langøyene Formation has dimensions of depth and lateral extend unlikely to have been formed by tidal currents, even though tidal channels may have been eroded as deep as several tens of meters in special geomorphological settings as straits and very large estuaries (Dalrymple, 2010; Martinius and van den Berg, 2011); (3) facies of sandstone and conglomerate above the unconformity surfaces is in favour of a fluvial, not a tidal origin. The shelf deposits of the epicontinental sea may lack tidal signals due to reworking by storms. However, some remains of bidirectional currents should have been recorded within the incised channel fills, at least in the finer-grained sandstone units.

The data presented in this study have been interpreted in terms of fluvial erosion, caused by fall in relative sea level, which was the base level of the system (Chapter 8.2). Fall in relative sea level can be caused by eustatic fall in sea level or from tectonic uplifts of the basin (Posamentier et al., 1988). In case of the Hirnantian time in the Oslo Region, these two types of mechanisms of fall in sea level may be attributed to
Caledonian tectonics, glacioeustacy or a combination.

Spjeldnæs (1957) suggested a major transgressive event from the south, flooding emergent areas in the Oslo Region during the Late Ordovician, but ascribed the erosional unconformity with overlying conglomerate beds in top of the Langøyene Formation as caused by Caledonian folding. Bjørlykke (1983) and Baarli (1990) suggested that Caledonian tectonics might have caused these sea-level fluctuations, whereas Jaanusson (1973); Bjørlykke (1974a) and Brenchley and Newall (1980) suggested that glacioeustacy associated with the Late Ordovician glaciation could have been the cause for sea-level fall and rise.

The Hirnantian glaciation on Gondwana (cf. Chapter 2.1) has been widely observed in the stratigraphic record in many parts of the World, with two major glacial phases (e.g. Sutcliffe et al., 2000; Ghienne, 2003). The two major erosional unconformities in the Hirnantian Langøyene Formation in the study area may correspond to these major glacial phases, by the simple reasoning that major glacial events in Gondwana (cf. Chapter 10.4.3), including fall in sea level in the order of up to 130 m (Kröger et al., 2015) should also be reflected in the shallow-marine succession in the Oslo Region.

The sedimentary analysis of the Langøyene Formation has shown that the succession has been deposited under overall shallowing conditions with significant sea-level fluctuations, accompanied by phases of deep erosion. Sandbakken (2014) suggested in his Master Thesis an incised valley system, showing at least three incisions, of which only one can be traced on Hovedøya. However, the present study shows that the third erosional phase was of much less significance than the previous two, which were correlated through the whole study area. The occurrence of erosional unconformities, their correlation and the occurrence of overlying conglomerate units has caused a series of misunderstandings in interpretations of the Langøyene Formation and subsequent discussions of causes for the conglomerate units.

The two phases of prominent sea-level lowstand represent a minimum sea-level drop of 65 to 70 m (Fig. 8.4, discussed in Chapter 8.2). Due to the short duration of the Hirnantian glaciation (between 0.5 to 1 Ma (Brenchley et al., 1994, 2003), see also Chapter 2.1), the sea-level must have been approximately the same before and after the glacial event, since this time span can be considered as too short to have yield sea-level changes caused by plate tectonic or local to regional crustal movements.
10.2. Incised valley model

This study shows that there have been several major erosional phases, connected to sea-level changes of at least 60 m, forming erosion as deep as 30 m into the sedimentary succession, relative to the datum plane. This estimated depth and the lateral extension of the erosional unconformities characterize the erosional feature as an incised valley complex (cf. Chapter 8), formed in response to glacioeustatic sea-level changes.

Chapters 6 and 7.2 outline that the first infill of the incised valley is characterized by fluvial conglomerate and sandstone deposits. From the facies distribution (Chapter 6), it appears that the fluvial streams of the lowstand systems tract at Rambergøya were of higher energy than in the streams of the corresponding units on Langøyene.

The infill successions of the valley system on Hovedøya are characterised by sediment bypass in channels cut down in the carbonate-platform in the upper part of the Langøyene Formation, leaving behind limestone blocks that were not transported. From these observations the high- and lower-energy fluvial stream conditions can be interpreted to represent proximal and distal fluvial environments, respectively. The Hovedøya and Rambergøya sites, with their locally derived huge clasts, may have been located close to the knickpoint on the incising fluvial system.

The outcome of this is that the incisions must have started somewhere in the southwest of the study area, and with proceeding sea-level fall incision continued northward (corresponding to landward direction). The time of maximum lowstand was succeeded by rising sea and base level, allowing sediments to start to fill the incised valleys.

From these observations a multi-phased incised valley system on the shallow, epicontinental shelf on the Baltic craton can be inferred as a depositional model for the Hirnantian time, as has been described in Chapter 8. In this context, the significance of the stratigraphy and the sedimentary facies on Bleikøya may have been underestimated in previous studies. Brenchley and Newall (1975) excluded observations from this locality from their discussion and interpretation, since they assumed that the succession on this island had thickened tectonically during Caledonian folding and thrusting. However, during the present study, no obvious tectonic repetition of strata has been recorded on Bleikøya, and no marked erosional unconformities have been observed in the Langøyene Formation here. In the incised valley model proposed in this thesis, Bleikøya represents an interfluve area, continued from the southern locality on Hovedøya with karstic interfluve surface on top of the Langøyene Formation, thus dividing two incisions towards the east and west, respectively (cf. Figs. 8.1, 8.2).
Karstic surfaces may be expected as correlative surfaces to the sequence boundaries as in other localities, representing sea-level lowstands.

Within the regional context in the eastward trending shelf-ramp (Brenchley and Cocks, 1982), the south- or even southwestward-trending orientation of the incised valley complex is noteworthy. Fluvial systems in foreland basins may be affected by foreland bulges or foredeeps, which can deflect their orientation. Plint and Wadsworth (2003) described an incised valley system, which has been followed the depression of the foredeep, parallel to the orogen. A similar scenario may be possible for the Oslo Region, where Bjørlykke (1983) and Baarli (1990) suggested that effects of foreland bulges may have had caused major sea-level changes. However, the studied succession, being allochthonous within the frontal detachment part of the Osen-Røa Nappe Complex (Nystuen, 1981, 1983, 1987; Morley, 1986; Bruton et al., 2010), formed closer to the Caledonian mountain belt in Late Ordovician time. The development of a possible foreland bulge in front of the southeastward moving Caledonian orogenic belt may have affected the direction of incised valleys, forced by glacioeustatic fall in sea level. A southwestward orientation of the incised valley complex might indicate that the incision took place on a very broad foreland bulge with an axis running parallel to a thrust-front that progressed towards the southeast.

### 10.3. Carbon isotope records

According to Delabroye and Vecoli (2010), the HICE (Hirnantian Isotope Curve Excursion) seems to be the most suitable tool for the correlation of shallow-water carbonates from Baltica and eastern Laurentia. The HICE presented in the present study for the central Oslo Region has been recorded from the succession on western Rambergøya (cf. Chapter 9.1, Fig. 7.1).

Another $\delta^{13}C$ curve from the central Oslo Region was published by Bergström et al. (2006) from Hovedøya. They found the start of the HICE in the uppermost meters of the Husbergøya Formation, within the brown weathering sandstone and a relative sharp end at the top of the Langøyene Formation. However, most of the Langøyene Formation remained unmeasured and peaks of the excursion are unknown. Kaljo et al. (2004b) presented $\delta^{13}C$ results from Upper Ordovician to Lower Silurian successions in the Oslo Region. Their maximum $\delta^{13}C$ excursion has its peak at 5.9‰ and was measured at Konglungen, about 12 km west-southwest of the present study area.

The here presented $\delta^{13}C$ curve is based on isotopes from bulk samples throughout the Upper Ordovician Husbergøya and Langøyene Formations (Chapter 9.1). A posi-
tive excursion has been recorded, starting in the upper ten meters of the Husbergøya Formation and ending in the brown weathering sandstone (facies H, FA 7, see description in Chapter 6). This excursion has been interpreted to represent the HICE for the section on Rambergøya, extending over a 62 m thick section. The maximum peak has a value of 5%. The onset of the excursion and the relative sharp decrease of δ13C values at the top of the Langøyene Formation match the results of Bergström et al. (2006) from Hovedøya. Since measurements were acquired throughout the Husbergøya and Langøyene Formations, two main phases in the excursion were observed, divided by a short decrease of δ13C values.

Kump et al. (1999); Bergström et al. (2006); Melchin and Holmden (2006) and Delabroye and Vecoli (2010) described Hirnantian δ13C curves based on studies in Nevada, Missouri-Illinois and Canada and compared curves from these studies with corresponding analysis from the Hirnantian in Laurentia, Baltica, southern China and Gondwana to show two phases in the Hirnantian isotope excursion. These two phases may also be seen in the δ13C curve from Rambergøya: a lower one, which is rather expressed as a plateau and an upper one, which appears as a single peak. The two main peaks appear above the lowermost two sequence boundaries in the here presented δ13C curve, respectively.

The comparison of the δ13C curve and the developed sea-level curve for the Oslo Region (Fig. 10.1) shows that the δ13C curve actually predates the first relative sea-level lowstand (Fig. 10.1B). The second phase of relative sea-level lowstand is followed by the second peak in δ13C values or may be interpreted to appear simultaneously.

However, studies by Saltzman and Young (2005); Jones et al. (2011) and Harper et al. (2014) have shown that the maximum drop in sea-level predates the maximum excursions recorded in δ13C. This is also valid for the data from the Oslo Region, if only the maximum value of +5% (second peak above SB2) is taken into account, compared to the maximum drop in sea-level of at least 65 m (Fig. 10.1B, cf. Figs. 8.4, 9.1). In case of a glacioeustatic sea level fall, large shelf areas and carbonate platforms would have been subaerially exposed in the epicontinental sea represented by the Upper Ordovician succession in the Oslo Region, and subsequently been eroded. This would support the weathering hypothesis by Kump and Arthur (1999) and Kump et al. (1999). This hypothesis is based on the assumption that δ13C excursions occur in response to fall in relative sea level and accompanied exposure and weathering of carbonate platforms and the effect of this upon the carbon cycle (Kump and Arthur, 1999). It has to be taken into account that the three or two major sequence boundaries represent significant hiatuses, which in turn modify the appearance of the δ13C curve and may thereby effect
Figure 10.1: Comparison of the measured δ\textsuperscript{13}C values and the sea-level curve based on sedimentologic and sequence stratigraphic analysis.

the global correlation of the Hirnantian succession from the Oslo Region with other studies.

Biostratigraphic data have to prove (or not) the synchronicity of the isotopic events between shaly and carbonate successions (Delabroye and Vecoli, 2010). This is of high importance when comparing the data acquired in the present study with global equivalent studies. Since the Langøyene Formation is characterized by sandstone, interrupted by erosional unconformities with overlying conglomerate units, the widely used graptolite zones are barely applicable. Studies on conodonts and chitinozoans are yet to be done for this succession to have an independent and testing correlation tool. Cathodoluminescence analysis has to be carried out to check for diagenetic recrystallisation or recrystallisation due to penetrating meteoric waters through strata. Even though there are no data from cathodoluminescence analysis available for the considered localities, the δ\textsuperscript{13}C may be reliable since meteoric alteration during diagenesis would shift measured δ\textsuperscript{13}C isotope data in a negative direction (cf. Saltzman and Thomas (2012)) and while considering a positive excursion as the HICE, this would result in weakened positive excursion, which may be regionally or globally correlated. However, a cathodoluminescence analysis should be executed to clarify possible recrystallisation, which may have affected the carbon-fractionation.
10.4. Regional and global context

10.4.1. Oslo Region

The incised valley system formed during sea-level lowstand is oriented towards south/southwest on a ramp-setting that has been described to trend towards southeast (Brenchley and Cocks, 1982). Given that erosional channels may meander, the observed orientation may differ from the actual orientation. However, the bayhead delta above the third sequence boundary is oriented towards southwest. The southwestward turn differs from the expected eastward trend and may have been caused by Caledonian nappe loading from the north/northwest and accompanied tectonic responses, as discussed above.

There occur major faults and differences in deformation style throughout the Oslo Region and adjacent areas, which may prevent the direct correlation and reconstruction of depositional environments throughout the area (Braithwaite et al., 1995). However, there occur similarities regarding depositional environments, which indicate a common origin of facies and facies belts, even if the relative positions amongst the considered localities remains uncertain.

In the northernmost part of Oslo Region, the Mjøsa district, the uppermost Ordovician may be absent (Mjøsa Hiatus) (Braithwaite et al., 1995; Bergström et al., 2010), or is characterized by a karstic surface indicating subaerial exposure (Braithwaite et al., 1995). As in the Langøyene Formation, the sediments below this hiatus are characterized by a shallowing upwards trend, which ended in this area with a subaerial emergence (Braithwaite et al., 1995).

Between the Mjøsa district and Oslo lays the Hadeland area. There, the Langøyene Formation corresponds with the Klinkenberg Formation (Braithwaite et al., 1995). This formation is characterized by a mixed carbonate-siliciclastic succession. Phases of subaerial exposure have been recorded as karstic surfaces on top of the Klinkenberg and the underlying Kalvsjøen formations, respectively (Heath and Owen, 1991; Braithwaite et al., 1995). Braithwaite and Heath (1992) presented a debris flow sedimentary model for the Klinkenberg Formation. The interpretation of possible debris flow deposits as being infills of incised channels, formed during a drop in relative sea-level, associated with adjacent subaerial exposed surfaces may be a reasonable re-interpretation of these settings. Further investigations under new perspectives in this area are, however, needed, in order to improve correlation of strata in the Oslo Region and compose comprehensive depositional models.

The absence of Upper Ordovician to Lower Silurian strata in the Ringerike district
(Spjeldnæs, 1957) fit well into the model of extensive subaerial exposure of shelf areas in the north and northwest of the central Oslo Region.

For large-scale reconstructions, it has to be kept in mind that the allochthonous character of the folded and thrust part of the Early Palaeozoic succession in the Oslo Region is not fully understood (Nystuen, 1987; Braithwaite et al., 1995; Bergström et al., 2010; Bruton et al., 2010). Investigations of conodont faunas from the Mjøsa area showed a more Laurentian than Baltic affinity (Bergström et al., 2010), may be indicating that the Upper Ordovician carbonate platforms of the area were originally located close to Laurentia.

10.4.2. Baltica

Ainsaar et al. (2010) suggested a Baltoscandian chemostratigraphy (Baltic Carbon Isotope Zonation) based on $\delta^{13}$C in combination with biostratigraphic data. According to their definition, the present study would be expected to show isotopic values that represent their zones BC16 and BC17. The former is defined to represent the steep increase of isotopic values during the Early Hirnantian and the latter to be the gently falling limb of the HICE towards the end of the Hirnantian (Ainsaar et al., 2010). The boundary between these two zones has been defined by the usage of chitinozoan biostratigraphy. Since studies on microbiostratigraphy in the central Oslo Region are yet to be done (see discussion above), fossil data cannot be used for Baltoscandian correlation, and the boundary between BC16 and BC17 in the Oslo Region thus remains unknown.

There have two major phases of erosion, connected to glacioeustatic sea-level lowstand been described by Bergström et al. (2006). They are referred to HA and HB in the following sections and discussed in Chapter 10.4.3, Laurentia.

Sweden

Lots of studies have been carried out in Upper Ordovician strata of central Sweden, e.g. in the Siljan district (e.g. Schmitz and Bergström, 2007; Ebbestad et al., 2015; Kröger et al., 2015). Schmitz and Bergström (2007) found two phases of major sea-level lowstand, referred to as HA and HB as earlier described by Bergström et al. (2006) on shallow-water succession in North America (see discussion below). Their findings correspond very well to the study by Kröger et al. (2015), who also described two phases of sea-level lowstand, based on investigations of speleothem in the Boda Limestone Formation. Ebbestad et al. (2015) presented a number of $\delta^{13}$C curves based on carbon-
ate rock bulk analysis for the Siljan district. They applied the Baltic Carbon Isotope Zonation by Ainsaar et al. (2010) and it appears that in most cases that where they detected the HICE, the BC16 extends from the onset of the HICE to the second major peak. If this should be the case, it may be suggested that the here presented $\delta^{13}$C curve from Rambergoya shows the boundary between BC16 and BC17 32 m above the base of the Langøyene Formation (42 m above the onset of the HICE). The phases of relative sea-level lowstand HA and HB, as they have been defined by Bergström et al. (2006), occur according to Ebbestad et al. (2015) within or above BC17. The present study would rather suggest that the major sea-level lowstands and the two-partition of the HICE occur simultaneously. However, this underlines the importance of a high-resolution biostratigraphic control in the Oslo Region to compare and fit the data with other studies in the Baltic Region.

Other studies on Hirnantian successions have been carried out in southern Sweden, in Västergötland (Loka Formation) (Bergström et al., 2006; Schmitz and Bergström, 2007). The Loka Formation has been described to consist of shallow-water, regressive sediments (Bergström and Bergström, 1996). The phases of relative sea-level lowstand (HA and HB) are marked by a hiatus, respectively, and have been identified at the lower boundary of the formation and on top of the middle member (Schmitz and Bergström, 2007, Fig. 5). This results in that both HA and HB occur within the HICE and not, as in most of the other localities presented in Schmitz and Bergström (2007), after the main peak. Comparative studies of the Loka Formation in Västergötland are recommended but due to little available data, correlations of both sedimentologic and isotopic data remain uncertain.

**Baltic states**

Investigations on Hirnantian successions have been carried out e.g. in Estonia and Latvia (e.g. Kaljo et al., 2001, 2004a; Hints et al., 2010). The Hirnantian started with a short sea-level rise and stabilization of the conditions in Latvia (lower boundary of the Kuldiga Formation) (Hints et al., 2010). This finding corresponds very well to the flooding surface identified at the lower boundary of the Langøyene Formation. There has been recorded a regressive sequence followed by a considerable hiatus in northern Estonia during the Hirnantian (Kaljo et al., 2001). The Hirnantian successions in southern Estonia appear to be more complete, but still they are interrupted by two erosional phases (Kaljo et al., 2001). They consist of two phases that show a transgressive development and turn into regression and (partial) subaerial exposure and erosion (Kaljo et al., 2001).
Reconstructions of changes in sea-level showed that the Hirnantian started with relatively high sea-levels, followed by two relative sea-level lowstands (Kaljo et al., 2004a). The first of these lowstands may be correlated to the erosive boundary between Kuldiga and Saldus Formation and was subsequently referred to as HA sensu Bergström et al. (2006) (Hints et al., 2010). The second erosive unconformity at the top of the Saldus Formation (Kaljo et al., 2001) has been referred to as HB (Hints et al., 2010). It is striking that ooid-containing sandstones have been described at the top of the Saldus Formation in southern Estonia (Kaljo et al., 2001), which may be correlated with the ooides in the FA 5, described previously.

The subaerial exposure and accompanied hiatus in Estonia appears to correspond to the Mjøsa hiatus in the northern Oslo Region. The southern Estonian shallow-water carbonates (Kaljo et al., 2001) appear to correspond to the facies found in the central Oslo Region. The position of HB in the Oslo Region compared to other localities on Baltica is not clear based on available data. However, the first significant fall in sea level in the Oslo Region (Fig. 8.4B) appears to correspond well to HA.

### 10.4.3. Global equivalents

**Laurentia**

Bergström et al. (2006) found two major stratigraphic gaps in the Hirnantian successions in Missouri and Illinois (Girardeau Limestone and Leemon Formation). They have been interpreted to represent phases of sea-level lowstands corresponding to the major periods of glaciations on Gondwana and named HA and HB (Bergström et al., 2006, Fig. 8). The first phase of sea-level lowstand, HA, was recorded in the middle Hirnantian (Bergström et al., 2006). This finding corresponds very well to the first drop in sea-level recorded in the Oslo Region (Fig. 8.4, stage B). The timing of HB appears to be somewhat different to the second relative sea-level lowstand observed in the Oslo Region (Fig. 8.4, stage D), given that the boundary between Ordovician and Silurian is set at the lower boundary of the Solvik Formation. Bergström et al. (2006) described that the phase named HB lasted until the middle Silurian, which is different to the observations in the Oslo Region, but may be explained by a diachronic character of the transgressive event connected to the end of the Hirnantian glaciation.

Investigations on sections from Anticosti Island (Canada) showed that this low-latitude locality in the west of the developing Caledonian Orogen, was affected by at least three major drops in relative sea-level (Ellis Bay Formation), recorded by tempestite frequency analysis (Long, 2007). Long (2007) argued that strata in the Baltic
Region must be missing since Brenchley and Newall (1980) and Kaljo et al. (2004a) described two sea-level drops. However, based on the present study from the Oslo Region it can be stated that it is possible to record three sea-level drops, but the third of them may be missing in some places due subaerial exposure at the time of the regression.

**Gondwana**

The Hirnantian glaciation (cf. Chapter 2.1) took place in extensive areas in western Gondwana (Scotese et al., 1999), and therefore the successions deposited through that time are expected to look very different than the ones deposited in low-latitude settings as they have been subject of this study. However, some results of studies carried out in West Africa (e.g. Ghienne, 2003) may be compared to the findings of the present study.

Ice-sheet advances and retreats are expected to correspond to global sea-level changes. Ghienne (2003) described a phase of glacial advance and a phase of deglaciation. The glacial maximum hereby has been reached by the maximum advance of the first glacial phase, and is followed by two minor glacial advances during a general deglaciation (Ghienne, 2003, Fig. 14). The subsequently presented diagram by Ghienne (2003) shows a remarkable similarity to the sea-level curve derived from the present study.
11 Conclusion

• The Upper Ordovician (Hirnantian) Langøyene Formation is cut by a major incised valley system on the islands Hovedøya, Rambergøya and Langøyene in the Oslo Region, formed in response to glacioeustatic fall during the Hirnantian glaciation on Gondwana.

• The incised valley system is defined by a lower unconformity and two other unconformities within an incised valley infill succession. The three unconformities are defined as sequence boundaries.

• The oolithic limestone in the top of the Langøyene Formation may be correlated with corresponding lithologies in Estonia and Latvia. A karstic surface on top of the limestone on Hovedøya represents an emerged interfluve area adjacent to the incised valley system.

• The incised valley system formed during glacioeustatic fall in sea level during two major phases, separated by sea-level rise and sediment infill. The second incision has cut through the first one on Hovedøya.

• Individual incision unconformities are at the base succeeded by fluvial conglomerate and sandstone beds, deposited by violently flowing streams, then followed by marine sandstone and shale in estuarine infill successions, including storm- and wave deposited strata and a bayhead delta.

• The orientation of the incised valleys was approximately north-south. Their origin may have been influenced by foreland bulging connected to the developing Caledonian orogen.

• The maximum glacioeustatic fall has been calculated to have been about 65 m.

• Carbon isotope analysis of bulk carbonate rock samples from Rambergøya detected the HICE starting in the upper 10 m of the Husbergøya Formation, continuing throughout the Langøyene Formation over a section of 62 m. The maximum $\delta^{13}$C peak has a value of +5 % and occurs above the second sequence boundary (SB2). $\delta^{13}$C values from Langøyene appear to be generally lower than the values measured on Rambergøya.
• The HICE recorded within the Husbergøya and Langøyene formations shows similar trends to carbon isotope data from other Hirnantian successions.


Allison, P. A. and Wright, V. P. (2005). Switching off the carbonate factory: A-tidality,


**A Supplementary figures**

**Figure A.1.** Complete log correlation of the Langøyene Formation on Hovedøya. The pink dot and box in the small map indicate the position of the photograph and the log at southern Hovedøya.
Figure A.2: Scaled sketch of the contact of the first and second major erosional unconformities on Hovedøya. The sketch represents the same area as shown in Figure 7.7. The conglomerate above the first sequence boundary SB1 can be easily identified, as well as the onlying parallel beds.
B Logs from Rambergøya, west
C Logs from Rambergøya, east
mainly bioclastic (50%) but also clasts

"deepening" surface succeeded by sandstone layers

contorted, locally collapse structures strongly developed

- A F S units, amygdaloidal parallel horizontal
D Logs from Langøyene, west
| m | LITH | mm | 0.00 - 0.125 | 0.125 - 0.25 | 0.25 - 0.5 | 0.5 - 1 | 1 - 2 | 2 - 3 | 3 - 4 | 4 - 5 | 5 - 10 | 10 - 20 | 20 - 40 | 40 - 80 | 80 - 160 | 160 - 320 | 320 - 640 | 640 - 1280 | 1280 - 2560 | 2560 - 5120 | 5120 - 10240 | 10240 - 20480 | 20480 - 40960 | 40960 - 81920 | 81920 - 163840 | 163840 - 327680 |
| 33 | coarsest | CLAY | SILT | VF | F | M | L | VC | G | PEB | 1 | 2 | 3 | 4 | 5 | 6 | 7 | 8 | 9 | 10 | 11 | 12 | 13 |
| 32 | coarsest | CLAY | SILT | VF | F | M | L | VC | G | PEB | 1 | 2 | 3 | 4 | 5 | 6 | 7 | 8 | 9 | 10 | 11 | 12 | 13 |
| 31 | coarsest | CLAY | SILT | VF | F | M | L | VC | G | PEB | 1 | 2 | 3 | 4 | 5 | 6 | 7 | 8 | 9 | 10 | 11 | 12 | 13 |
| 30 | coarsest | CLAY | SILT | VF | F | M | L | VC | G | PEB | 1 | 2 | 3 | 4 | 5 | 6 | 7 | 8 | 9 | 10 | 11 | 12 | 13 |
| 29 | coarsest | CLAY | SILT | VF | F | M | L | VC | G | PEB | 1 | 2 | 3 | 4 | 5 | 6 | 7 | 8 | 9 | 10 | 11 | 12 | 13 |
| 28 | coarsest | CLAY | SILT | VF | F | M | L | VC | G | PEB | 1 | 2 | 3 | 4 | 5 | 6 | 7 | 8 | 9 | 10 | 11 | 12 | 13 |
| 27 | coarsest | CLAY | SILT | VF | F | M | L | VC | G | PEB | 1 | 2 | 3 | 4 | 5 | 6 | 7 | 8 | 9 | 10 | 11 | 12 | 13 |
| 26 | coarsest | CLAY | SILT | VF | F | M | L | VC | G | PEB | 1 | 2 | 3 | 4 | 5 | 6 | 7 | 8 | 9 | 10 | 11 | 12 | 13 |
| 25 | coarsest | CLAY | SILT | VF | F | M | L | VC | G | PEB | 1 | 2 | 3 | 4 | 5 | 6 | 7 | 8 | 9 | 10 | 11 | 12 | 13 |
| 24 | coarsest | CLAY | SILT | VF | F | M | L | VC | G | PEB | 1 | 2 | 3 | 4 | 5 | 6 | 7 | 8 | 9 | 10 | 11 | 12 | 13 |
| 23 | coarsest | CLAY | SILT | VF | F | M | L | VC | G | PEB | 1 | 2 | 3 | 4 | 5 | 6 | 7 | 8 | 9 | 10 | 11 | 12 | 13 |
| 22 | coarsest | CLAY | SILT | VF | F | M | L | VC | G | PEB | 1 | 2 | 3 | 4 | 5 | 6 | 7 | 8 | 9 | 10 | 11 | 12 | 13 |
| 21 | coarsest | CLAY | SILT | VF | F | M | L | VC | G | PEB | 1 | 2 | 3 | 4 | 5 | 6 | 7 | 8 | 9 | 10 | 11 | 12 | 13 |
| 20 | coarsest | CLAY | SILT | VF | F | M | L | VC | G | PEB | 1 | 2 | 3 | 4 | 5 | 6 | 7 | 8 | 9 | 10 | 11 | 12 | 13 |
| 19 | coarsest | CLAY | SILT | VF | F | M | L | VC | G | PEB | 1 | 2 | 3 | 4 | 5 | 6 | 7 | 8 | 9 | 10 | 11 | 12 | 13 |
| 18 | coarsest | CLAY | SILT | VF | F | M | L | VC | G | PEB | 1 | 2 | 3 | 4 | 5 | 6 | 7 | 8 | 9 | 10 | 11 | 12 | 13 |
| 17 | coarsest | CLAY | SILT | VF | F | M | L | VC | G | PEB | 1 | 2 | 3 | 4 | 5 | 6 | 7 | 8 | 9 | 10 | 11 | 12 | 13 |
| 16 | coarsest | CLAY | SILT | VF | F | M | L | VC | G | PEB | 1 | 2 | 3 | 4 | 5 | 6 | 7 | 8 | 9 | 10 | 11 | 12 | 13 |
| 15 | coarsest | CLAY | SILT | VF | F | M | L | VC | G | PEB | 1 | 2 | 3 | 4 | 5 | 6 | 7 | 8 | 9 | 10 | 11 | 12 | 13 |
| 14 | coarsest | CLAY | SILT | VF | F | M | L | VC | G | PEB | 1 | 2 | 3 | 4 | 5 | 6 | 7 | 8 | 9 | 10 | 11 | 12 | 13 |
| 13 | coarsest | CLAY | SILT | VF | F | M | L | VC | G | PEB | 1 | 2 | 3 | 4 | 5 | 6 | 7 | 8 | 9 | 10 | 11 | 12 | 13 |
| 12 | coarsest | CLAY | SILT | VF | F | M | L | VC | G | PEB | 1 | 2 | 3 | 4 | 5 | 6 | 7 | 8 | 9 | 10 | 11 | 12 | 13 |
| 11 | coarsest | CLAY | SILT | VF | F | M | L | VC | G | PEB | 1 | 2 | 3 | 4 | 5 | 6 | 7 | 8 | 9 | 10 | 11 | 12 | 13 |
| 10 | coarsest | CLAY | SILT | VF | F | M | L | VC | G | PEB | 1 | 2 | 3 | 4 | 5 | 6 | 7 | 8 | 9 | 10 | 11 | 12 | 13 |
| 9 | coarsest | CLAY | SILT | VF | F | M | L | VC | G | PEB | 1 | 2 | 3 | 4 | 5 | 6 | 7 | 8 | 9 | 10 | 11 | 12 | 13 |
| 8 | coarsest | CLAY | SILT | VF | F | M | L | VC | G | PEB | 1 | 2 | 3 | 4 | 5 | 6 | 7 | 8 | 9 | 10 | 11 | 12 | 13 |

**GEOLOGIST:** F. FRANECKE
**FORMATION:** LANGØYENE TH.
**AGE:** LATE ORDOVICIAN
E Logs from Langøyene, east
F Logs from Hovedøya
### Log 1, 4/4

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<th>SHEET</th>
<th>DATE</th>
<th>SCALE</th>
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<td>63.10.44</td>
<td>1:50</td>
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</table>

#### GEOLOGIST: Transcendence
#### FORMATION: Liquefier
#### AGE: Late Eonization

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</table>

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- To be continued in a second log.
- Looks like the end of the island.
- Sedimentary due to the fact that the samples are normally not exposed.
- Sandstone gets rougher near the middle and becomes more coarse near the top.
- Scattered bedding structures seen in picture.
- Lentic lense on a gradual scale.
- Slowly rising though to a hard, undisturbed and undamaged layer.
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<tr>
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<td>44.5</td>
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**LOG 2, 1/1**

**LOCALITY**

**LITH**

**SAND**

**VC**

**G**

**FEB.**

**DATE:** 05/10/14

**SCALE:** 1:50

**GEOLOGIST:** FranziskaTRANNECK

**FORMATION:** LAMINATED

**AGE:** LATE ODOMNIAN

---

Note: The diagram includes various geological observations and measurements, but the text is not fully transcribed in this format. Additional details and annotations are also present on the page.
<table>
<thead>
<tr>
<th>m</th>
<th>LITH</th>
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</table>

**LOCALITY:**
- **Houdagya Springs - Sandhills**

**GEOLOGIST:**
- **J. A. Terlouw**

**FORMATION:**
- **LARGO GREYWACE**

**AGE:**
- **LATE DINONIAN**

- Deep troughs fanned out, badly very weathered
- Angularized fragments
- All layers apparent, almost in the background, at least the conformations
- Coarse sand, with 98%

**Notes:**
- Rocks up to 10 cm, mostly argillaceous
dust, occasional of tough, banded at 0.7 cm, high
99.7% crinoid, some undetermined for 1 mm in
LARGO OFF and the formation patterns
- About 3 blocks
### Log 6, 1/2

**Locality:** S. Borehole [5]

**Sheet:** 1

**Geologist:** J. Z. Franek

**Formation:** Lancelin Group

**Age:** Late Oligocene

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<th>Depth (m)</th>
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<tr>
<td>0 - 1.5</td>
<td>Clay 0.5%</td>
</tr>
<tr>
<td>1.5 - 2.5</td>
<td>Clay 1%</td>
</tr>
<tr>
<td>2.5 - 3.5</td>
<td>Clay 1.5%</td>
</tr>
<tr>
<td>3.5 - 4.5</td>
<td>Clay 2%</td>
</tr>
<tr>
<td>4.5 - 5.5</td>
<td>Clay 2.5%</td>
</tr>
<tr>
<td>5.5 - 6.5</td>
<td>Clay 3%</td>
</tr>
<tr>
<td>6.5 - 7.5</td>
<td>Clay 3.5%</td>
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<td>7.5 - 8.5</td>
<td>Clay 4%</td>
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<tr>
<td>8.5 - 9.5</td>
<td>Clay 4.5%</td>
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<tr>
<td>9.5 - 10.5</td>
<td>Sand 10%</td>
</tr>
</tbody>
</table>

**Observations:**
- Fragments of coral, annelid, bivalves, foraminifera
- Porous structure
- Fractures present
- Bedding irregular

**Remarks:**
- Bedding is well developed
- No significant changes in lithology

**Scale:** 1:50
<table>
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<th>Layer</th>
<th>Lithology</th>
<th>Description</th>
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<tbody>
<tr>
<td>C1 ( coarse-grained sandstone)</td>
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<td>Well sorted, medium to coarse-grained sandstone with cross-bedding and ripple marks.</td>
</tr>
<tr>
<td>C2 (coarse-grained sandstone)</td>
<td></td>
<td>Similar to C1 but with more angular grains.</td>
</tr>
</tbody>
</table>

Note: All data (except 0.5m below) - accurate, through cross-bedded set.

- 0.5m: Mixed sequence, includes fine-grained siltstone and clay layers.
Hovedøya south

LOCALITY: Hovedøya

LITH:

mm 1 2 3 4 5 6 7 8 9 10 11 12 13 14 15 16 17 18 19 20 21 22 23 24 25 26 27 28 29 30 31 32 33 34 35 36 37 38 39 40 41 42 43 44 45 46 47 48 49 50 51 52 53 54

CLAY: VF F M C VC PEB

SILT: 0.00-0.05

G: 0.1

F: 0.05-0.1

M: 0.5

C: 1

VC: 10

PEB: 1

AGE: HIRDASTIAN

FORMATION: LANGøyENE

GEOLOGIST: F. TRANGE

SCALE: 1:50

DATE: 21.09.14

Sheet A of A

NOTES:
- rough: x-beded set
- no distinct transport direction
- Furrows locally dunes, westward transport direction
- Westward accretion - dunes ~10 cm 2
- rough x-beded set
- Horizontal parallel laminated set
G List of samples
### Table G.1.: List of samples taken on Rambergøya, the position in section is given in metres, relative to the lower boundary of the Langøyene Formation.

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<th>Sample</th>
<th>Position in the section [m]</th>
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<td>Rw20</td>
<td>18</td>
</tr>
<tr>
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### Table G.2.: List of samples taken on Langøyene, the position in section is given in metres, relative to the lower boundary of the Langøyene Formation.

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<td>50,5</td>
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<td>Lw14</td>
<td>55</td>
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## H Dataset of isotopic analysis

Table H.1.: Stable isotope measurements from Rambergøya.

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<td>Rw P 06</td>
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<td>Rw P 09</td>
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### APPENDIX H. DATASET OF ISOTOPIC ANALYSIS

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<td>Rw P 29</td>
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### I Dataset of XRF analysis

**Table I.1.:** XRF measurements and errors for Titanium, Zirconium and Rubidium for the samples from Rambergøya west.

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### APPENDIX I. DATASET OF XRF ANALYSIS

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**Table I.2.** XRF measurements and errors for Calcium and Strontium for the samples from Rambergøya west.

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